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SOME BASIC PARAMETERS ASSOCIATED WITH THE FLOOD RAINS AT CHICAGO, OCTOBER 9–12, 1954

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ABSTRACT

The development of heavy rains during the period October 9–12, 1954, which led to record floods in the Chicago area, is investigated. An attempt is made to ascertain whether the low-level convergence and vertical motion could be accounted for by terms in the vorticity equation associated with vorticity advection, thermal advection, and stability. Moisture transport into the area is also analyzed.

1. INTRODUCTION

The present investigation was begun as part of a program for studying convergence in the lower troposphere and associated precipitation patterns. The program has been conducted jointly by the Weather Forecasting Research Center at the University of Chicago and the U. S. Weather Bureau Forecast Center in Chicago.

The period of heavy rainfall during October 9–12, 1954 is significant not only because of the record floods in the Chicago area, and the great amount of property damage [1], but also because the largest (for that area) recorded 48-hour accumulations of rainfall occurred. From the synoptic meteorologists' point of view the situation was especially interesting because the thundershowers occurred mostly in the warm air south of a front which was in the general vicinity during this period, and the mechanism for releasing the instability was not immediately clear.

The excessive amounts of rainfall appeared to be confined almost entirely to extreme northern Illinois as depicted in an isohyetal map (fig. 1) of official Weather Bureau reports and unofficial reports collected by the Illinois State Water Survey [2].

For a more complete synoptic description of this series of showers the reader is referred to the paper by Nash and Chamberlain [3]. Two sea-level charts from their study are reproduced in figures 2 and 3 for synoptic orientation.

The purpose of this paper is to examine quantitatively several basic factors related to the rainfall occurrence. An attempt is made to explore in some detail the occurrence of low-level convergence as implied by some of the parameters suggested by Petterssen's [4] development equation and to describe certain features of the low-level moisture jet.

2. APPLICABILITY OF TERMS IN THE DEVELOPMENT EQUATION

The development equation (Petterssen [4]) defines development in terms of low-level convergence or vertical motion, and, since vertical motion and convergence tend to produce stability changes, a study of the applicability of the development equation to this heavy thundershower situation has seemed appropriate.

This has been done by attempting to analyze data for the Chicago flood situation in the various functional forms of the separate terms of the equation:

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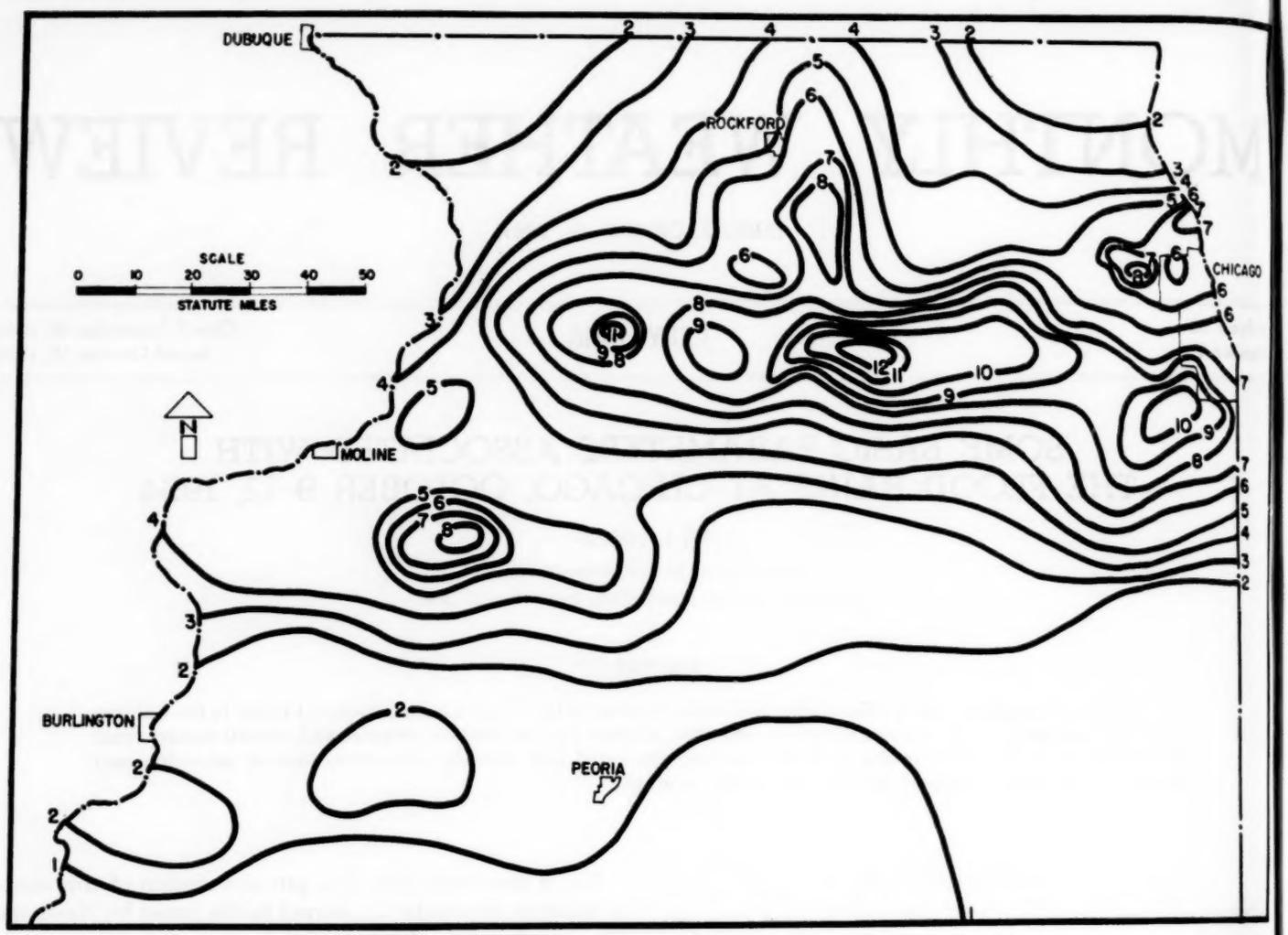

FIGURE 1.—Total storm rainfall (inches) October 9–10, 1954. From [2]).

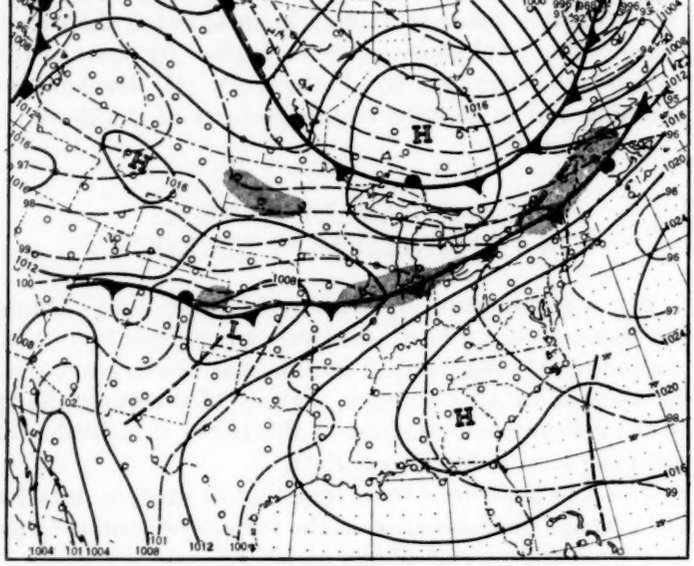

FIGURE 2.—Surface chart at 0330 GMT on October 10, 1954 and 1000–700-mb. thickness contours in hundreds of feet (dashed) at 0300 GMT on October 10, 1954. Shading shows active precipitation. (From [3]).

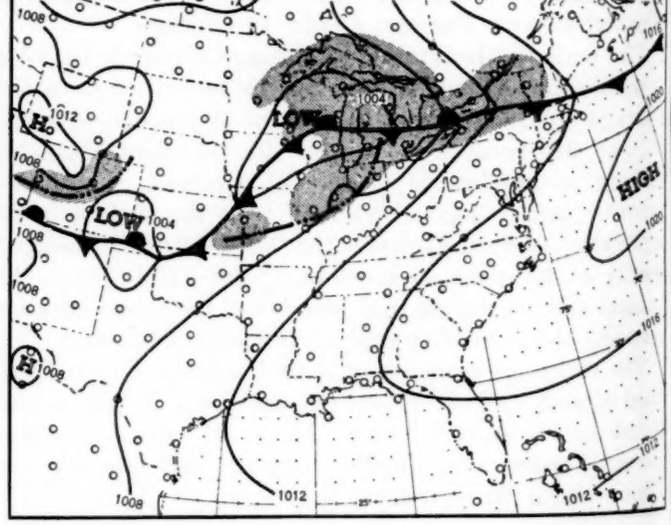

FIGURE 3.—Surface chart at 0330 GMT on October 11, 1954. (From [3]).

$$-Q_0D_0=A_{QL}+\mathbf{V}_0\cdot\nabla Q_0-\frac{g}{f}\nabla^2A_h$$

$$-\frac{R}{f}\nabla^2\left[\log\frac{p_0}{p}\left(\overline{\omega(\Gamma_a-\Gamma)}+\frac{1}{c_p}\frac{\overline{dW}}{dt}\right)\right]$$

$-Q_0D_0$ represents the product of absolute vorticity and development (convergence) at sea level or 1000 mb.

A_{QL} represents $-(\mathbf{V}\cdot\nabla Q)_L$ or vorticity advection at the level of non-divergence.

A_h represents $-\overline{\mathbf{V}}\cdot\nabla h$ or advection of an isobaric layer whose thickness is h.

p is the pressure at the level of non-divergence.

$\overline{\omega(\Gamma_a-\Gamma)}$: $\omega=\frac{dp}{dt}$ denotes vertical velocity where p is pressure and t is time.

$\Gamma_a=\frac{dT}{dp}$=adiabatic lapse rate of temperature T with respect to pressure.

$\Gamma=\frac{\delta T}{\delta p}$=actual lapse rate with respect to pressure.

The bar denotes the mean value from the level of non-divergence to 1000 mb.

$\frac{1}{c_p}\frac{\overline{dW}}{dt}$: c_p is specific heat at constant pressure.

$\frac{\overline{dW}}{dt}$ is the heat (other than latent) supplied to a unit mass per unit time.

R is the gas constant.

Subscript 0 denotes sea level (or 1000-mb.) value.

Vorticity advection patterns in the vicinity of, and upstream (southwest quadrant) from, northern Illinois at sea-level (A_{Q_0}) and in the middle (500 mb.) troposphere or near the tropopause (300 mb. or 200 mb.) (A_{QL}) did not reveal any large contribution to low-level convergence. From an examination of the charts for October 9, 1500 GMT, October 10, 0300 GMT and October 10, 1500 GMT, the largest value of A_{QL} was found to be 5.8×10^{-9} sec.$^{-2}$. This was computed from the 200-mb. chart for October 10, 1500 GMT.

No attempt was made in this limited study to locate the height of the level of non-divergence. It was assumed to be at 300 mb. or above, which experience (Petterssen, [4]) has shown to be frequently the case before intensification of any sea level system, and in this case no intensification of any sea level system of the scale of extratropical cyclones occurred.

However, analysis of patterns of the second term, the Laplacian of thermal advection, revealed larger contributions for the layer 1000–500 mb. as shown in figures 4 to 6 amounting to 30×10^{-9} sec.$^{-2}$ in the vicinity of or upstream (southwest quadrant) from northern Illinois. These data were computed separately using a 300-km. grid for layers 1000–700 mb. and 700–500 mb. and added algebraically. Neither the 500–300-mb. nor 500–200-mb. layer was used in these computations because the thermal advection was relatively weak for those layers in this situation.

Two pronounced minima of the Laplacian of thermal advection are revealed in figures 4 and 6 as approaching northern Illinois on the 9th and 10th respectively. These two patterns suggested the occurrence of two specific areas of low-level convergence, while an intermediate pattern of opposite sign, figure 5, contributed to low-level divergence. The two patterns of low-level convergence appeared to be associated in time and space with: (1) the beginning of showers on the afternoon of the 9th (the first indication in the regular reporting network of Weather Bureau stations was a trace of rain at Bradford, Ill., on the 1830 GMT chart); and (2) the development of heavy showers which moved across the Chicago area on the afternoon of the 10th.

The stability, or rather potential instability ($\overline{\Gamma_a-\Gamma}$) was evaluated from a Showalter-type stability index [5], but using values from that level at or below 700 mb. having the *largest equivalent potential temperature* (usually the level of greatest moisture content), rather than 850-mb. values, unless of course that level had the largest equivalent potential temperature. The difference in the 500-mb. temperature of the ambient air and the temperature at 500 mb. of the lifted parcel (lifted from the level of maximum equivalent potential temperature) gives a measure of the difference between adiabatic lapse rate and actual lapse rate, $(\Gamma_a-\Gamma)$, for a given layer. The temperature and pressure for the parcel to be lifted and that for the ambient air are initially the same since they are taken initially at the same significant level. Values of this index of potential instability were plotted and analyzed. The fact that lapse rates were taken over slightly different height intervals (but most were from near 900 to 500 mb.) did not influence the values of the index as much as would the selection of one arbitrary level as being most representative for maximum lifted parcel temperature values at 500 mb.

It is notable in this case that successive soundings showed that lapse rates consistent with conditional instability were maintained despite the usual tendency for lapse rates to approach the pseudo-adiabatic in cases of heavy rain. In this case the potentially unstable air continued to flow into the more localized convergence area and the heavy precipitation was not sufficiently widespread to modify the lapse rates of this potentially unstable air upstream.

The stability-vertical motion term from the development equation,

$$-\frac{R}{f}\nabla^2\log\frac{p_0}{p}\overline{\omega(\Gamma_a-\Gamma)}$$

cannot be quantitatively evaluated since no measurements of vertical motion are available for this situation. However, the area of upward motion can be qualitatively outlined making use of indirect aerology through the

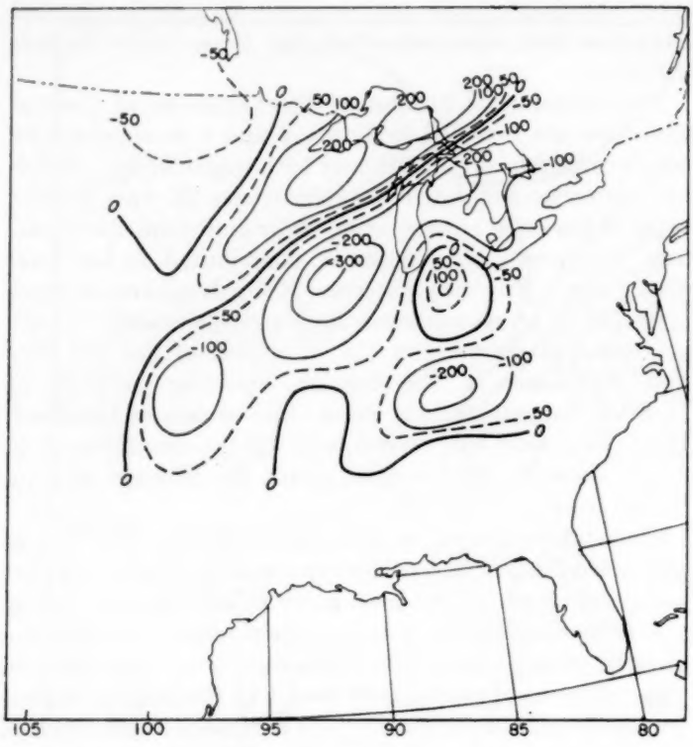

FIGURE 4.—Laplacian of thermal advection for 1000–500-mb. layer, October 9, 1954, 1500 GMT (units$\times 10^{-10}$ sec. $^{-2}$).

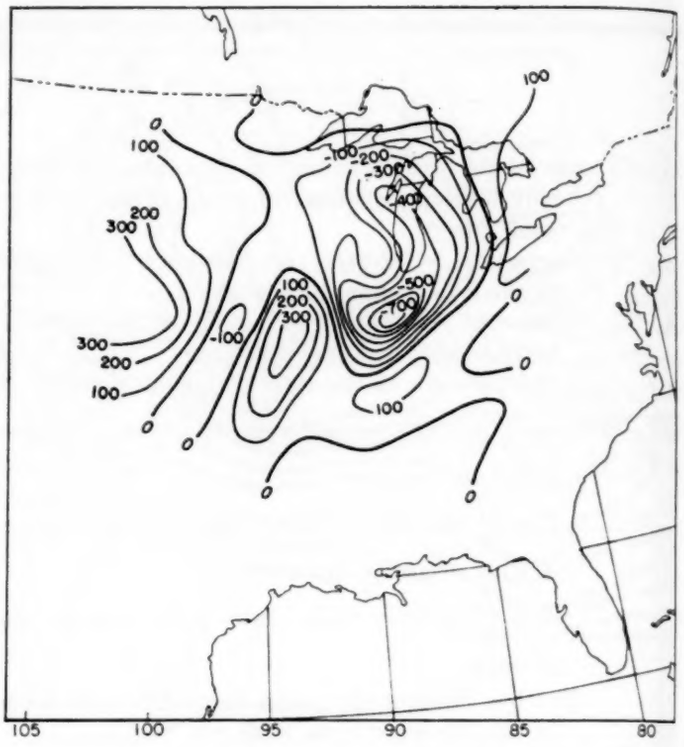

FIGURE 6.—Laplacian of thermal advection for 1000–500-mb. layer, October 10, 1954, 1500 GMT.

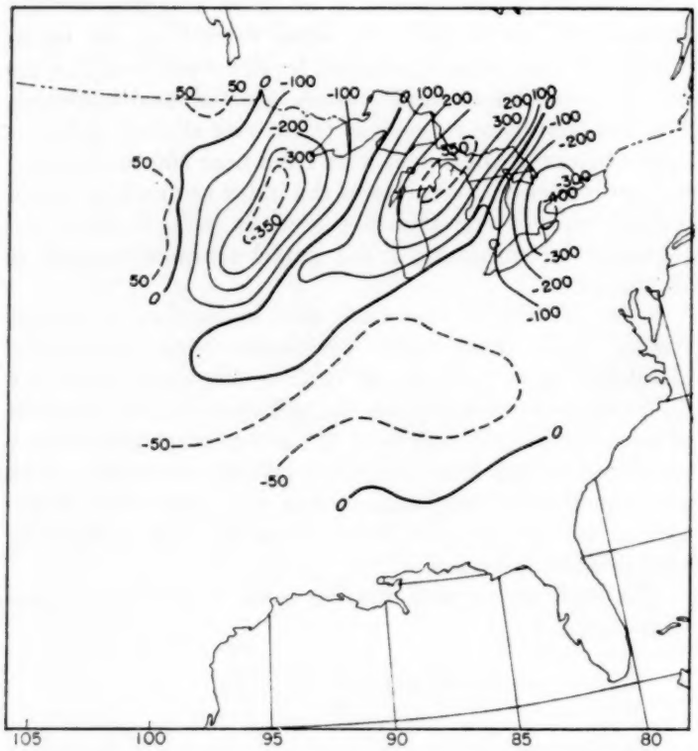

FIGURE 5.—Laplacian of thermal advection for 1000–500-mb. layer, October 10, 1954, 0300 GMT.

examination of cloud data. Overcast areas are cross-hatched on the patterns of the stability factor in figures 7–11. Reference to these figures then gives a qualitative indication of the contribution of the Laplacian of this term. These areas of potential instability over or just upstream from Chicago are consistent with the development and movement of low-level convergence and precipitation over the Chicago area on the afternoons of the 9th and 10th, and with the continuance of some precipitation into the early morning hours of the 10th.

A discussion of the 12-hour continuity of these patterns before and during the period of heavy rain follows:

At 1500 GMT on the 8th (fig. 7) more than 24 hours before the rains began, considerable cloudiness was associated with a trough northwest of Chicago, and some light rains occurred, mostly to the north of Chicago. However, analyses of the lapse rates showed that important potential instability was not yet indicated by the index of lapse rate difference which gave a plus 1 value at Omaha and Topeka. (Larger negative values of the index indicate greater instability).

By 0300 GMT on the 9th (fig. 8) a marked change in the potential instability factor values had occurred but the area of overcast skies remained north and east of Chicago suggesting that little vertical motion had developed upward from the lowest moist layer by this time, especially to the south and west of Chicago.

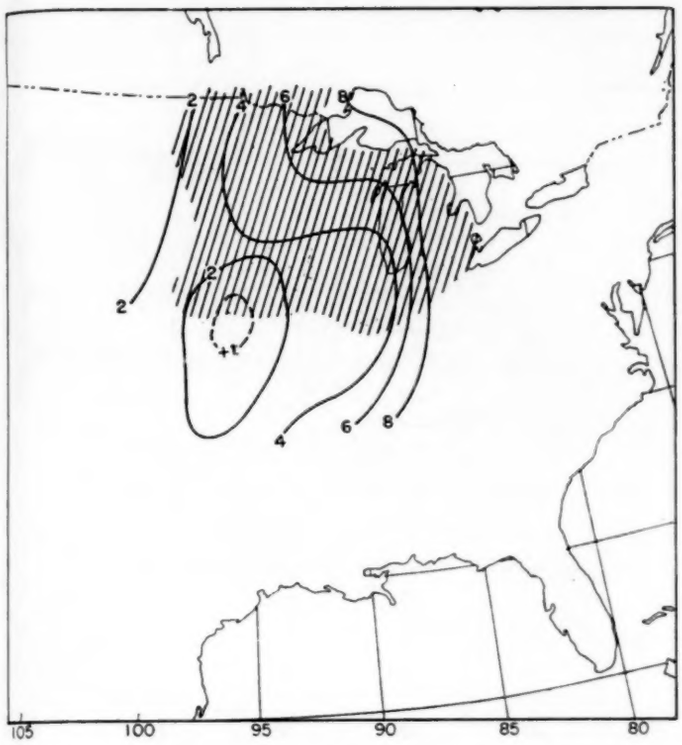

FIGURE 7.—Stability index, October 8, 1954, 1500 GMT. Cross-hatched area represents mostly overcast skies.

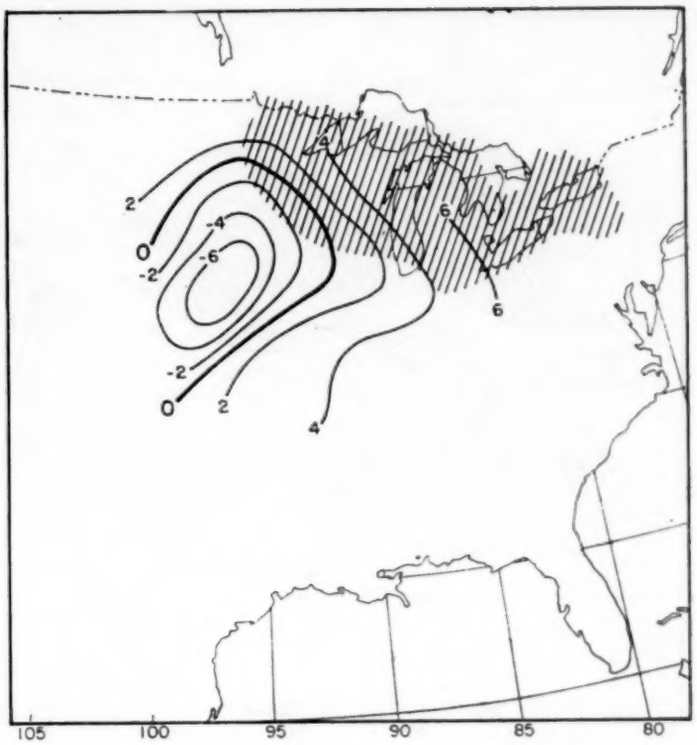

FIGURE 8.—Stability index, October 9, 1954, 0300 GMT. Cross-hatched area represents mostly overcast skies.

Not until the 1500 GMT chart of the 9th (fig. 9) was there a good indication of cloudiness developing upstream from Chicago within the area of potential instability and even then a Laplacian computation using a small grid (assuming uniform upward motion over the entire cloudy area) would have given little immediate indication of any important contribution by the vertical motion-stability term to low-level vertical motion or convergence. This was due to rather straight isolines of the instability factor and also little shear in the pattern. However, it is recognized that such detail in the analysis is not very reliable due to wide spacing of the points for which data were available.

The chart of the 10th at 0300 GMT (fig. 10) showed a much larger curvature to the isolines of stability index from northwestern Illinois eastward to southeastern Lower Michigan. This pattern was within the area where considerable cloudiness was observed and upward motion was presumed to be present. With upward motion and sharp curvature of the stability index pattern in the proper sense, the Laplacian of vertical motion-stability term could make an appreciable contribution toward low-level convergence.

A computation from the center of the area of potential instability included within the cloudy area, using an assumed upward wind speed (uniform over the entire area) of 10^{-2} mb. sec. $^{-1}$ showed a value of 12×10^{-9} sec.$^{-2}$ as a

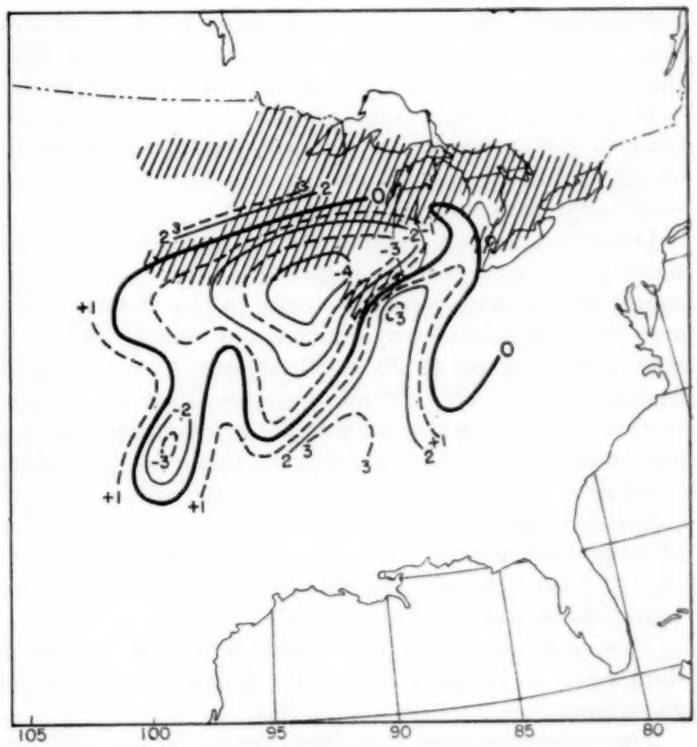

FIGURE 9.—Stability index, October 9, 1954, 1500 GMT. Cross-hatched area represents mostly overcast skies.

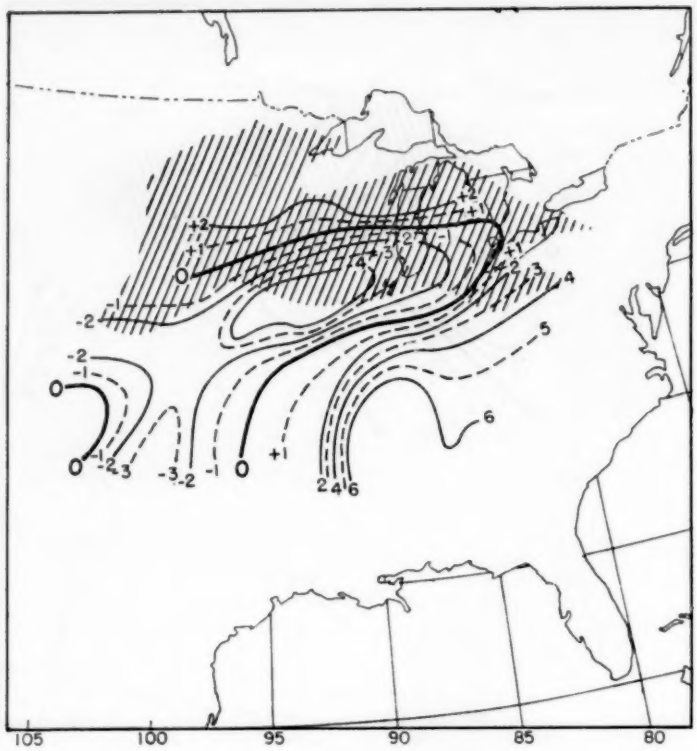

FIGURE 10.—Stability index, October 10, 1954, 0300 GMT. Cross-hatched area represents mostly overcast skies.

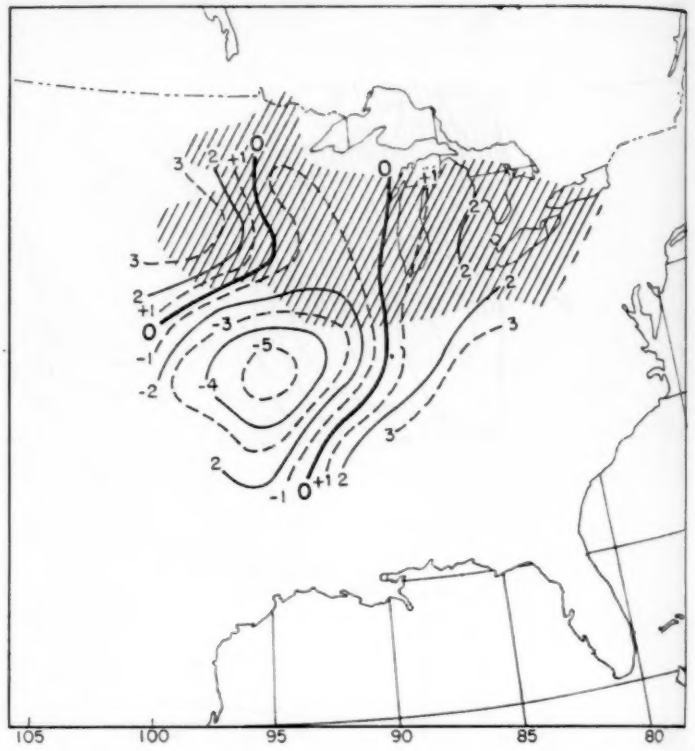

FIGURE 11.—Stability index, October 10, 1954, 1500 GMT. Cross-hatched area represents mostly overcast skies.

representative contribution of this term, which is greater than the vorticity advection computation but less than the Laplacian of thermal advection values in this series.

If upward motions were greater where the potential instability was also greater, as seems likely, the Laplacian of *the product* of vertical motion and instability in this term could have been somewhat larger.

Data from certain rainfall studies and from a recent study by Smagorinsky [6] suggest that the average horizontal gradients of vertical velocity in the vicinity of an area of heavy rainfall (say 5 inches and more in 24 hours) may easily amount to 20 to 80 cm. sec.$^{-1}$ over the grid distance of 300 km. When such gradients of vertical motion with a central maximum are superimposed upon a maximum of potential instability, then the individual instability factors multiplied by the individual vertical velocities leads to Laplacian values which are half an order of magnitude greater than the original computation of 12×10^{-9} sec.$^{-2}$ which was obtained using a uniform upward field of motion of about 10^{-2} mb. sec.$^{-1}$.

This additional magnitude would make the buoyancy term equal to or greater than the Laplacian of thermal advection.

The buoyancy term then becomes an increasingly important consideration in cases of this type as vertical velocities increase over more or less local areas giving a chain reaction or "bootstrap effect" contribution to low-level convergence.

Again on the 10th at 1500 GMT (fig. 11) a significant contribution was indicated by the Laplacian of the instability (assuming upward motion) with the cloudy area upstream (southwest quadrant) from northern Illinois.

No practicable means was available for evaluating the non-adiabatic term in the development equation.

The heavy rains indicated that low-level convergence was present and quite strong. But little surface development as evidenced by increasing cyclonic circulation and deepening of central pressures, occurred in this situation.

In an attempt at explanation one might conjecture that low-level vorticities (Q_0) apparently were localized into small patterns the size of groups of thunderstorm cells and in the absence of any strong large-scale vorticity advection aloft tended to dissipate perhaps due to friction and turbulent transport of momentum and vorticity near the ground and also possibly due to thermal modifications with cooling in the areas of heavy rain. The recurring strong thunderstorm activity and associated convection to high levels suggests that the level of non-divergence did not drop and that compensation occurred mainly between the stratosphere and troposphere rather than within the troposphere. The lack of surface circulation intensification on an extratropical scale may also be consistent with the lack of any marked vorticity advection at high levels, which according to the working hypothesis concerning development (Petterssen, Dunn, and Means [7]) is usually present with surface development.

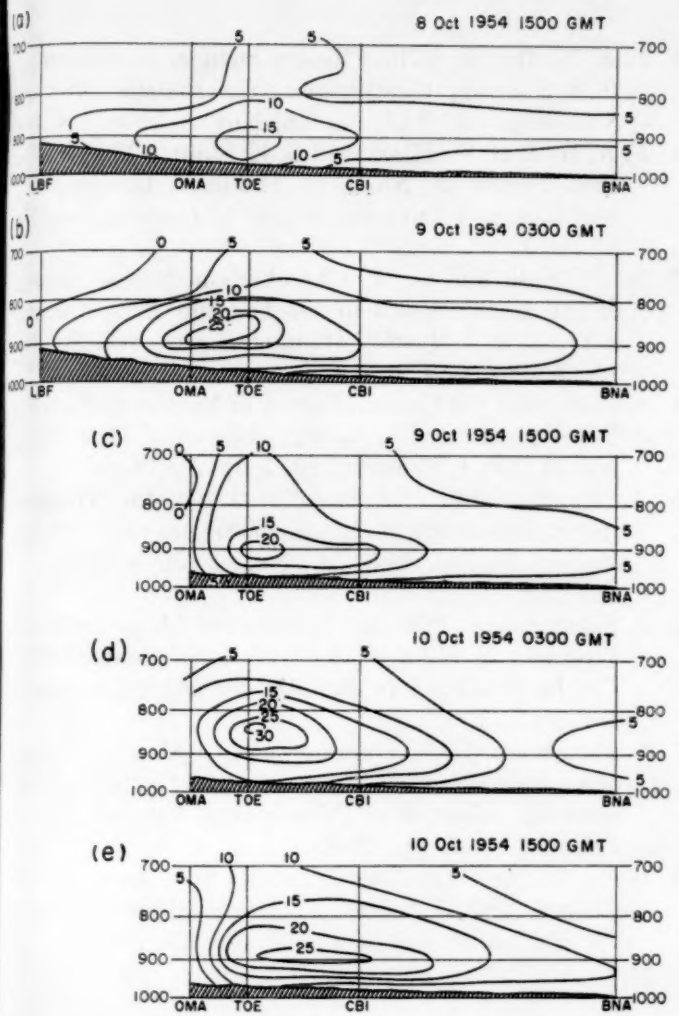

FIGURE 12.—Moisture transport normal to cross-section, October 8-10, 1954. Units in gr. sec.$^{-1}$ for a unit cross-section 1 mb. high by 1 cm. wide.

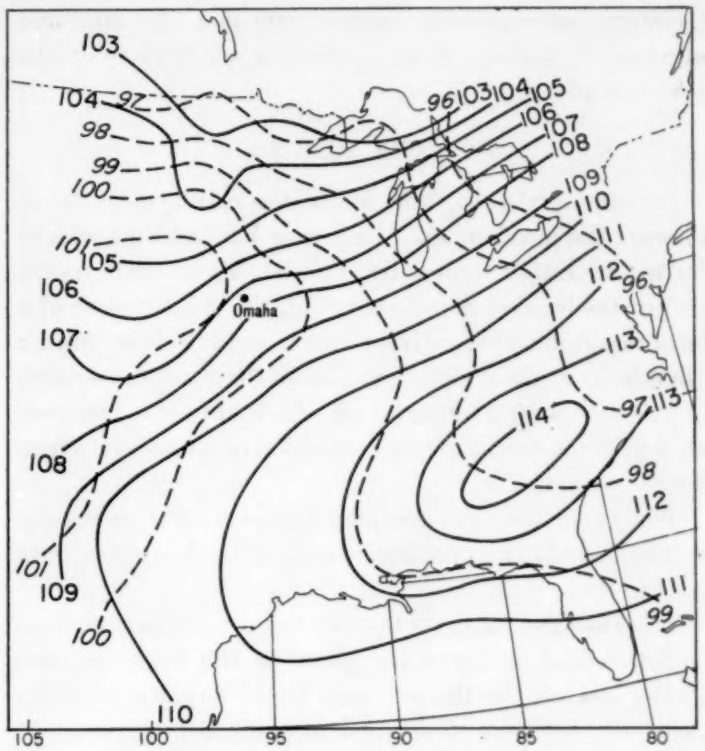

FIGURE 13.—Thickness and mean flow, 1000–700 mb. October 9, 1954, 0300 GMT. Dashed lines are thickness labeled in 100's of feet. Solid lines are mean flow labeled as sum of 1000- and 700-mb. heights in 100's of feet.

The data suggest that the Laplacian of thermal advection contributed significantly initially to the low-level vertical motion and convergence in these heavy rains, and that after the development of cloudiness and convective activity, the Laplacian of the vertical motion-stability term also may have contributed quite significantly to the maintenance of low-level convergence.

3. FORECASTING SIGNIFICANCE

The question arises from the forecasters' point of view as to the possibility of identifying the mechanism that was responsible for developing and increasing potential instability. Charts for two periods (October 8, 1500 GMT and October 9, 0300 GMT) are given in figures 7 and 8. The potential instability was not only present but strong with a minus 6 central value on the 9th at 0300 GMT as contrasted with a plus 1 central value at 1500 GMT, October 8, giving a net drop of 7° C. in the index of potential instability for the Omaha soundings over this 12-hour period. The marked change was associated with an influx of warm moist air below 800 mb. and especially near the 900-mb. level. This influx of warm moist air was associated with the development of the low-level jet (fig. 12 a and b). The 1000–700-mb. mean contour and thickness patterns upstream from Omaha (fig. 13) show the typical contour-isotherm relationship (Means [8]) with warmer air toward lower contour values, which is hydrostatically consistent with the occurrence of the low-level jet. The corresponding 850-mb. chart shows similar patterns.

The primary forecast difficulty in this case was due to the failure of appreciable cloudiness to develop upstream from Chicago until almost the very beginning of the period of heavy thundershowers, also the lack of any well-defined impulse as identified by a katallobar-anallobar couplet which forecasters usually associate with moving or developing weather situations.

The potential instability was maintained throughout the period of heavy rains by continued transport of warm moist air in the low-level jet. This is depicted in figure 12 for each of the 12-hour periods from the 8th at 1500 GMT through the 10th at 1500 GMT. The low-level moisture transport increased markedly during the period October 8, 1500 GMT to October 9, 0300 GMT and remained strong while the jet core moved very little laterally, remaining in the area between Columbia, Mo. and Omaha, Nebr.

Greatest transport was between 900 and 850 mb. and averaged 3 times as great in value in the core as at 700 mb. just above the jet core.

4. CONCLUSIONS

In summary then, the forecaster might attempt to forecast the development of a strong but localized area of warm advection especially in the lower troposphere (below the level of non-divergence), the development of a local minimum of negative stability index values, and an abundant supply of moisture. All of these were associated in this case with a low-level jet which in turn was associated with the formation and extension of a low-level warm tongue.

The chain reaction contribution of locally increasing vertical velocities to the importance of the buoyancy term is suggested.

This case also suggests the value of a maximum equivalent potential temperature point in the lower portion (below 700 mb.) of the sounding for a Showalter stability index since values of equivalent potential temperature taken at 850 mb. are not always representative of high values that are found below 700 mb., especially near the 900-mb. level.

The maximum moisture transport was found more frequently in the vicinity of the 900-mb. level than at 850 mb.

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WINDS AND PRESSURES OVER THE SEA IN THE HURRICANE OF SEPTEMBER 1938

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ABSTRACT

The winds and pressures over the sea in the New England hurricane of September 1938 are reconstructed, making the meteorological ingredients available for a meteorological-oceanographic investigation of the record tide produced by this storm on the southern New England coast. The methods of analysis of the meteorological data are applicable to other hurricanes.

1. INTRODUCTION

The hurricane of September 1938 was among the most intense known to have occurred along the Atlantic Seaboard. This rapidly-moving storm reached the coastline of New England at the time of high tide, and as Brooks [1] describes it: "Towering surges on this combined astronomical tide and storm wave threw the sea to such heights that demolition was general along the exposed coast, and they came so suddenly that hundreds of persons, some of them at the shore to watch the fine surf, were engulfed and drowned." The purpose of this study was to develop reasonable estimates of winds and pressures over the sea in this great storm in order that they might be correlated with the observed tides which are the highest of record over much of the southern New England coast. The study was conducted in cooperation with the Corps of Engineers and was initiated in connection with their investigations on possible protective works against hurricane tides at Narragansett Bay and other points on the southern New England coast. However, the results should also be of general interest to those investigating the problem of predicting the tides to be expected with hurricanes.

The basic data for this study consisted of barograph traces and autographic wind records for all land stations in the area from New England to Hatteras and ships' observation forms. Time-graphs of the sea level pressure and 10-minute averages of the wind direction and speed were prepared for each station. Original Weather Bureau maps and special studies of the storm (e. g. [1, 3, 12, 18]) have been used to guide the detailed examination of the storm. Since the storm occurred before the development of aircraft reconnaissance and radar for meteorological purposes, the analysis techniques are more indirect than those that can be used on more recent hurricanes.

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2. TRACK OF STORM CENTER

A detailed storm track (fig. 1), necessary as a reference point for the wind and pressure analyses, was taken from Pierce's maps [12] with certain modifications. The 1200 EST position was moved northward on the basis of more complete ship observations than were available to Pierce. The track was also altered slightly over New England in order to indicate the position of the pressure center only. For instance, the location of Pierce's center is closer to New Haven than to Hartford, although Hartford had the lowest pressure. The wind center did pass west of New Haven (the pressure center was to the east) which probably influenced his decision on the position of the storm center. In this study a separate wind center was found and its position relative to the pressure center determined by plotting the wind direction (relative to the pressure center) for both Hartford and New Haven every 15 minutes, or more often during rapid direction changes (fig. 2). The wind center was found to be 16 nautical miles southwest of the pressure center. It was assumed that this distance and direction between the centers would hold while the storm was over the ocean. This observed displacement of the wind and pressure centers has been compared with Shaw's [16] theoretical separation of the centers for a moving circular depression. The relationship for the displacement, d, given by Shaw is $d=\frac{C}{2\omega \sin \phi+\zeta}$, where C is the speed of movement of the storm, $2\omega\sin\phi$ the Coriolis parameter, and ζ the angular velocity of the storm. At 1500 EST the gradient wind was 96 m. p. h. at a distance of 50 nautical miles from the center (the radius of maximum wind), which gives an angular velocity of 46×10^{-5} rad. sec.$^{-1}$. If these values are used, the computed displacement is 18.7 nautical miles, which is in close agreement with the 16 measured from figure 2.

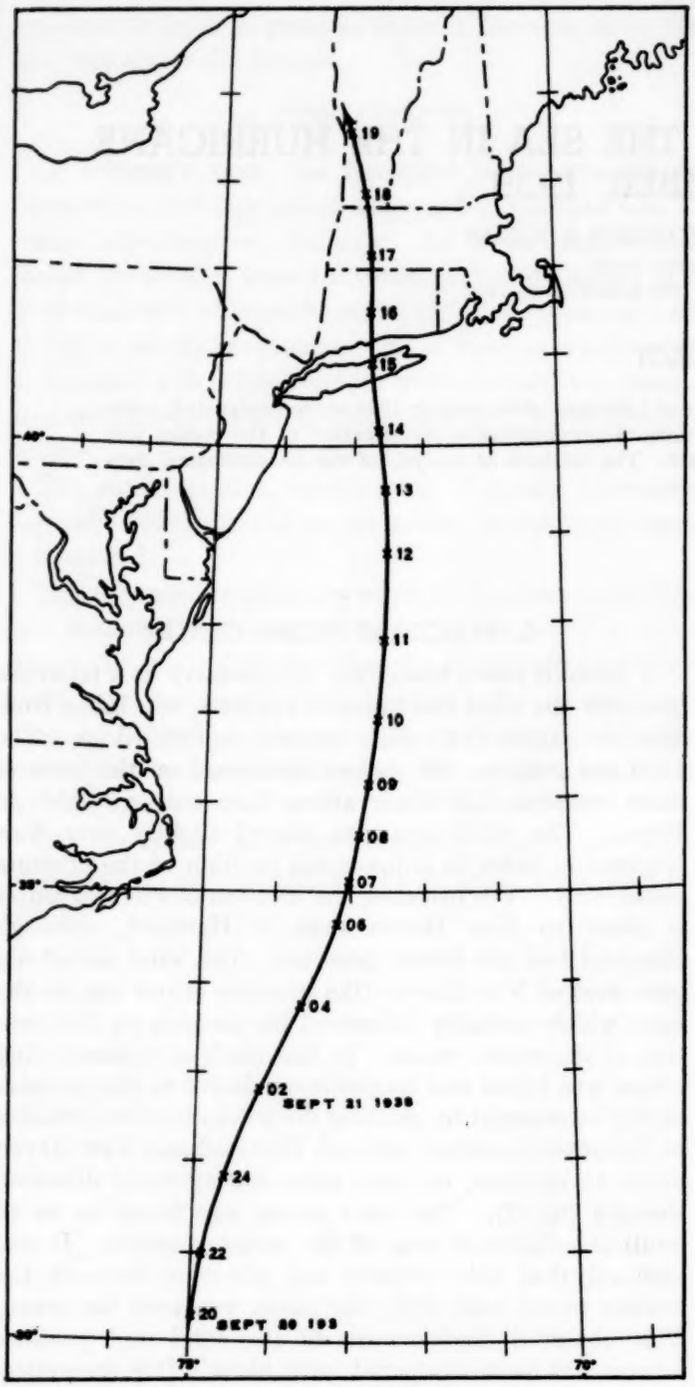

FIGURE 1.—Track of pressure center, hurricane of September 20–21 1938. Plotted times are in EST.

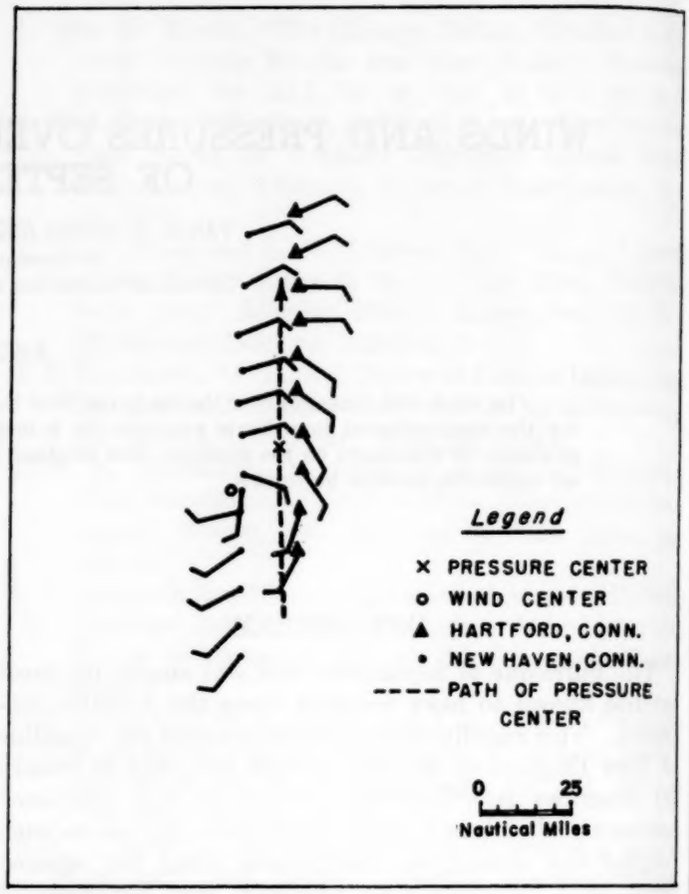

FIGURE 2.—Wind direction plotted relative to pressure center, near 1500 EST, September 21, 1938 at Hartford and New Haven, Conn. Pressure center (X) is reference point for plotting winds. Wind center (o) derived subjectively from plotted wind directions.

3. PRESSURE DISTRIBUTION

Pressure analyses were made hourly from 1200 EST through 1900 EST, and the maps for 1200, 1400, 1500, and 1900 EST are reproduced in figure 3. The pressure pattern was nearly circular through 1600 EST, but by 1900 EST it had become more elongated. Radial pressure profiles in the four cardinal directions were plotted from the maps for each hour, and the hourly continuity of these profiles was in turn used to adjust the analyses in areas of no data. Although the storm was over the ocean for the most part at 1200 and 1300 EST, ship reports taken to the north and west of the storm center at the 1200 EST standard observation time provide enough data for a fairly adequate analysis, together with continuity with later times when the storm was over land. Selected profiles along a line to the east of the pressure center, approximately normal to the direction of motion, are shown in figure 4; variation of the central pressure with time, derived from the profiles, is depicted in figure 5. The central-pressure determinations over land are considered reliable within a few hundredths of an inch, the estimates over the sea much less so, with the reliability more appropriately expressed in quarters of an inch. A central pressure of 27.75 inches at 1200 EST (fig. 5) is derived by extrapolating the pressure profile inward from the ship reports, of which 28.10 inches (corrected) by the *Birmingham City* was the lowest. The central-pressure curve in figure 5 is leveled off prior to 1200 EST, on the basis that the central pressure is also estimated at about 27.75 inches on the previous afternoon, with the storm center near 30° N. The lowest report at that time was 28.00 inches (whether corrected was not specified) from the *Indian Arrow*.

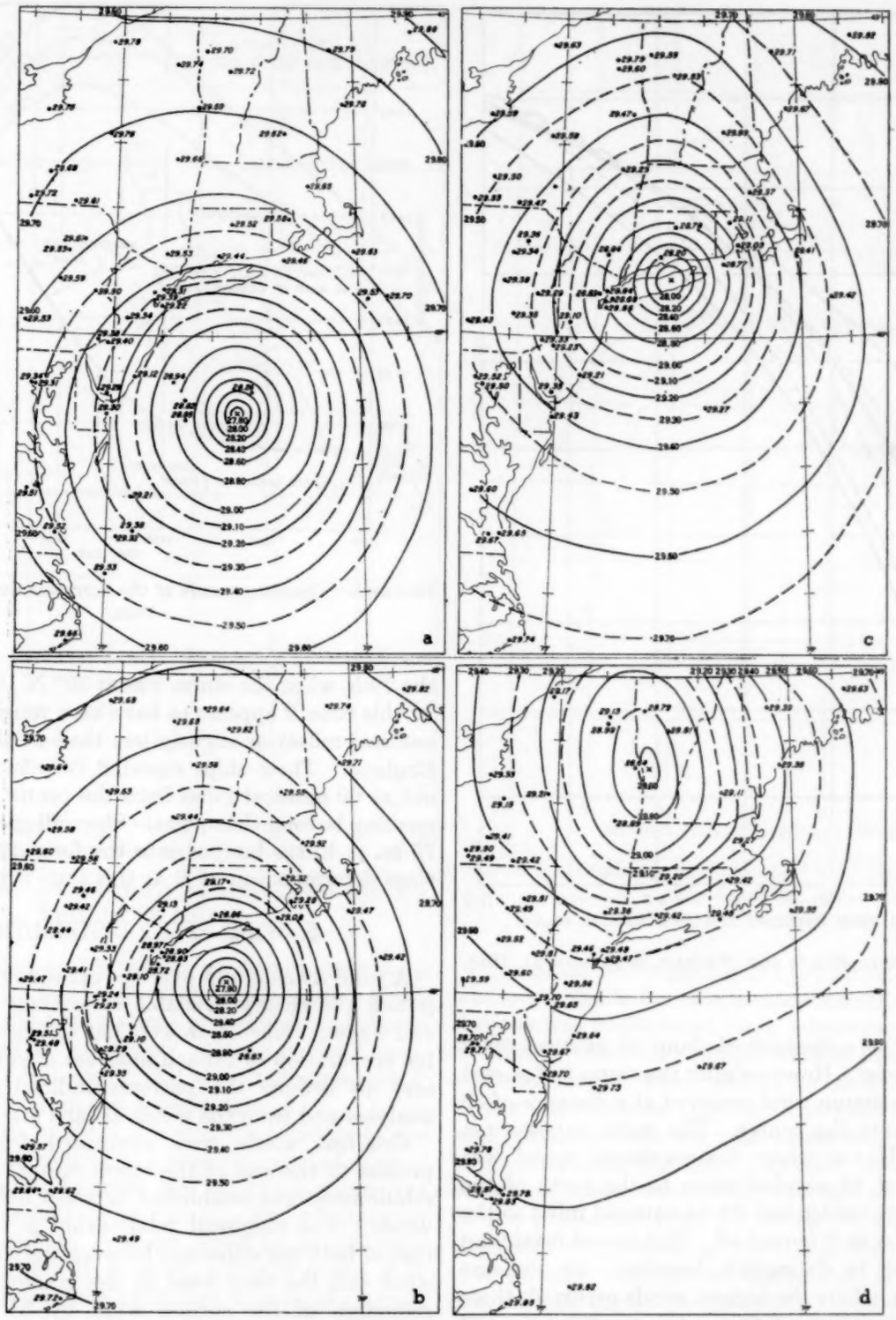

FIGURE 3.—Sea-level pressure (inches), September 21, 1938. (a) 1200 EST, (b) 1400 EST, (c) 1500 EST, (d) 1900 EST.

4. RADIUS OF MAXIMUM WINDS

Reid [13, 14] has developed a method for estimating the relative tide-producing potential of hurricanes over a particular part of the Continental Shelf as a function of the central pressure of the storm and of the radius to the maximum wind speed, R. The tide potential, by his method, is quite sensitive to R, principally because the larger the circle of maximum winds, the longer the fetch of strong winds in approximately the same direction. In the 1938 storm R is large and seems to vary somewhat around the storm. In the northern part of the storm, Hartford reached its maxi-

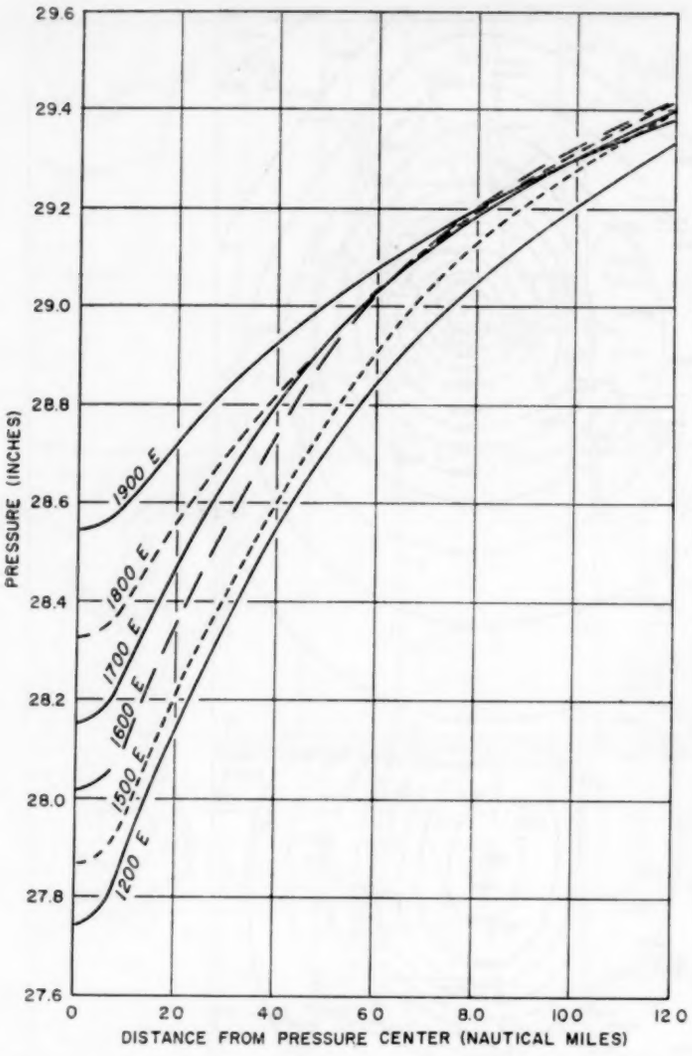

FIGURE 4.—Pressure profiles to east of center, September 21, 1938.

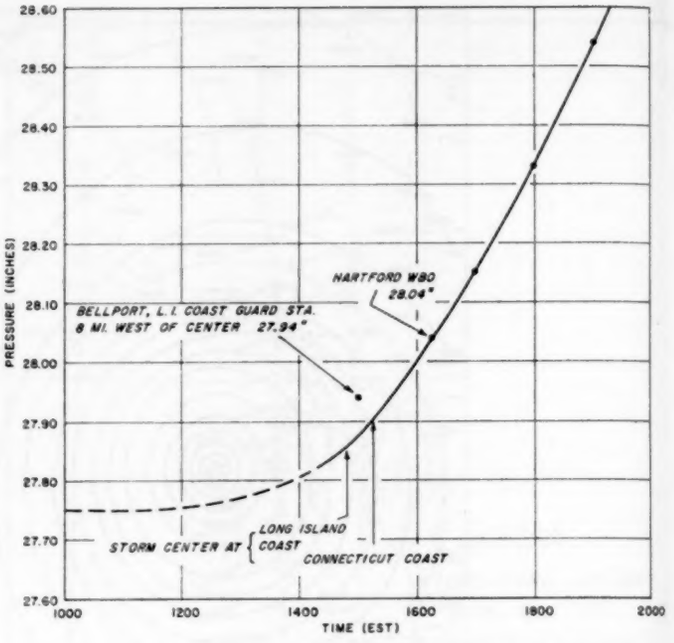

FIGURE 5.—Central pressure of the hurricane during September 21, 1938.

mum wind speed at a distance of about 50 nautical miles from the wind center. However after the storm had passed Hartford, the maximum wind occurred at a distance of 30 nautical miles from the center. The same pattern was shown at New Haven where the maximum speed was observed at about 43 nautical miles to the north of the approaching storm center and 30–35 nautical miles to the south of the center as it moved off. The second maximum was more difficult to distinguish, however. On the east side of the storm, where the highest winds occurred, there is an absence of stations between the wind center and about 60 nautical miles, and R was placed at the radius indicated by the computed gradient winds in this direction (figs. 6 and 7). Other evidence for a large R is a report of "calm" winds as far out as 30 nautical miles [1], and a report prepared by the Corps of Engineers [3] in which the region of strongest winds was estimated at a distance of 64 nautical miles to the right of the storm center. There were more ships reporting in the storm area on the afternoon of the 20th, when the storm was at 30° N., than on the 21st. At this time R appears to have been roughly the same (50 nautical miles) or slightly less than at the coast of New England. These ships reported Beaufort force 12 as far out as 60 nautical miles from the center, with speeds decreasing beyond that point. Since all speeds greater than 75 m. p. h. are lumped into the force 12 classification, a close determination of R at this time is not possible.

5. WIND SPEED DISTRIBUTION

A wind distribution over the ocean is required for computing the energy available to produce the storm surge and waves. Since data were not sufficient over the ocean for preparation of isotach analyses directly, it was necessary to deduce the patterns indirectly from pressure analyses and from the winds at land stations.

Gradient winds were computed from the pressure profiles to the east of the storm center, and an empirical relationship was established between these and the 30-ft. winds. The empirical relationship is intended to take care of both the difference between the computed gradient wind and the true wind at the gradient level [2] and the reduction of the surface wind by friction. The solid lines in figures 6 and 7 are the gradient wind curves. These curves for the hours 1400, 1500, and 1600 EST (fig. 6) are compared with the observed surface winds at Block Island, Providence, Nantucket, and New Haven. New Haven was actually north of the wind center, but three observations nearest the center were plotted on the

[2] Some of the assumptions used in computing the gradient wind which are not fulfilled in the hurricane are balance of forces, wind parallel to isobars, and radius of curvature of trajectory same as radius of curvature of isobar.

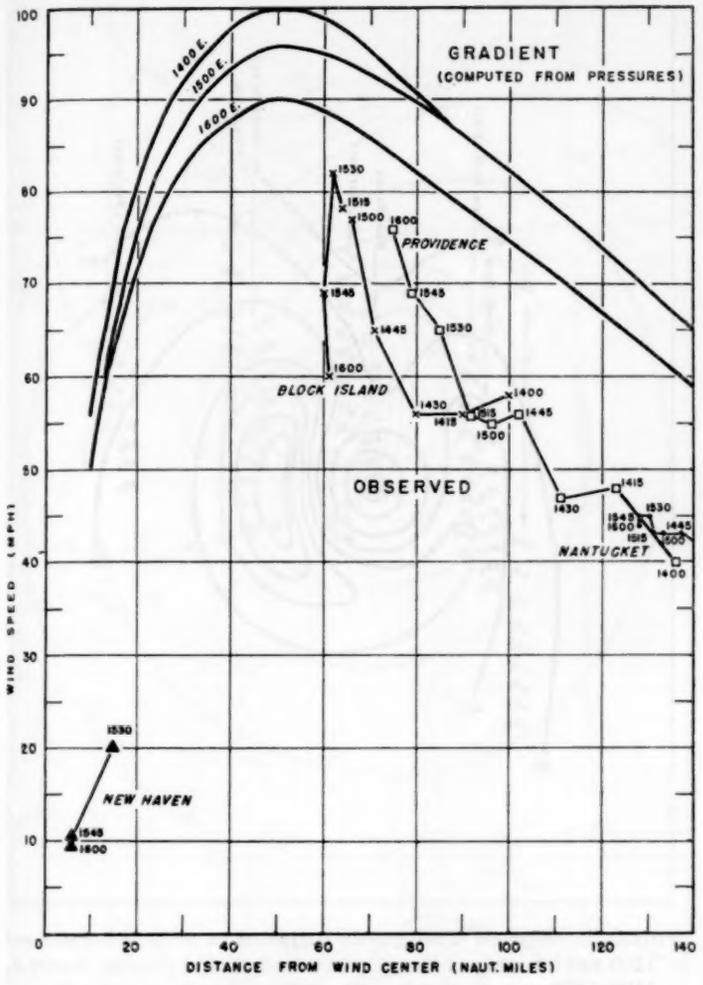

FIGURE 6.—Gradient and observed wind speeds, 1400–1600 EST, September 21, 1938.

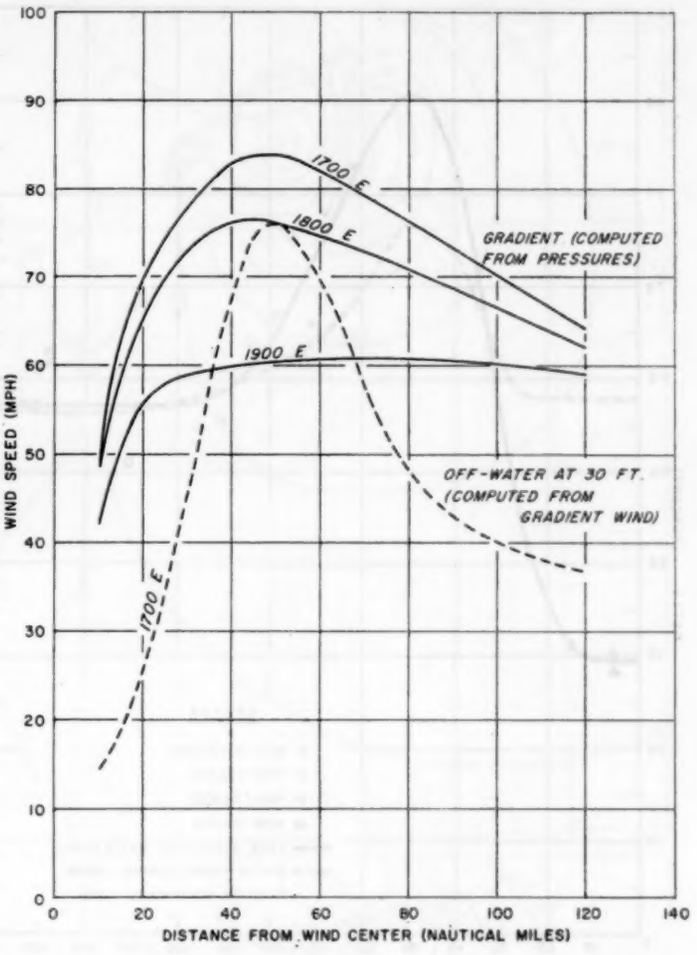

FIGURE 7.—Gradient and "off-water" wind speeds 1700–1900 EST, September 21, 1938.

assumption that the central values would not vary much around the storm. The height of the anemometers above ground varies at the four stations. The observed speeds were then adjusted to "off-water" (defined in the Appendix) values at 30 feet, using the wind reduction graph in figure 12 (explained in the Appendix).

The ratio of each of these adjusted wind speeds to the gradient wind speed at the same time was plotted against distance from the wind center (fig. 8), and a curve was drawn to fit the ratios [3] (solid curve). The relationship of the surface wind to gradient wind in the Florida hurricane of August 26–27, 1949 [8], adjusted to the R of the 1938 hurricane, is shown by dashed lines on the figure, and was used as a guide in drawing this curve. Profiles of the surface winds at 30 feet can be computed from the gradient-wind profiles at any time by using this relationship of the surface wind to the gradient wind. An example of such a profile is shown by the dashed curve of figure 7.

To obtain an overall isotach pattern to supplement the above profiles to the east, a composite map of wind observations was constructed for a 2-hour period centered at 1500 EST. Mean 10-minute wind speeds, adjusted to a common basis,[4] were plotted relative to the pressure center and analyzed (fig. 9). The wind center from this analysis agrees with the wind center determined from the wind directions (fig. 2). The 1500 EST profile of the surface off-water wind at 30 feet (not shown but similar to the profile for 1700 EST in fig. 7) was used to help determine the analysis to the east in the area of no data.

Over-water isotachs, the end-product of the wind-speed analysis, are shown for 1200, 1400, 1500, and 1600 EST in figure 10. Winds prior to 1200 EST may be estimated by transposing the pattern for 1200 EST along the track (fig. 1). Over-water isotachs for specified times can be derived from the composite pattern of figure 9 by increasing the speeds by 12 percent to adjust from off-water to over-water (see Appendix) and by applying an additional small adjustment for filling of the storm between the time con-

[3] The 1545 and 1600 EST wind speeds at Block Island were not considered representative and were not used to develop ratios. They were also disregarded in the analysis of figure 9. At this time the wind had veered to a direction from behind a hill higher than the anemometer.

[4] Adjusted to 30 feet above the surface and to the "off-water" frictional category by use of figure 12, and for filling of the storm by multiplying each speed by the ratio of the gradient wind at that time (from fig. 6) to the 1500 EST gradient wind.

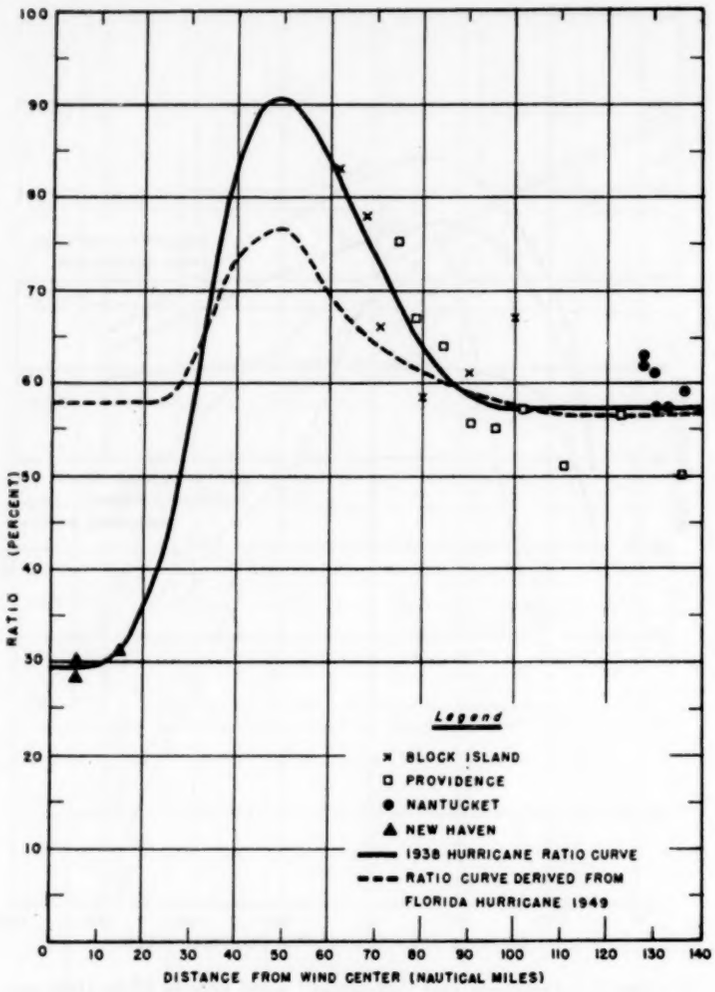

FIGURE 8.—Ratio of surface wind speed ("off-water" at 30 ft.) to gradient wind, September 21, 1938.

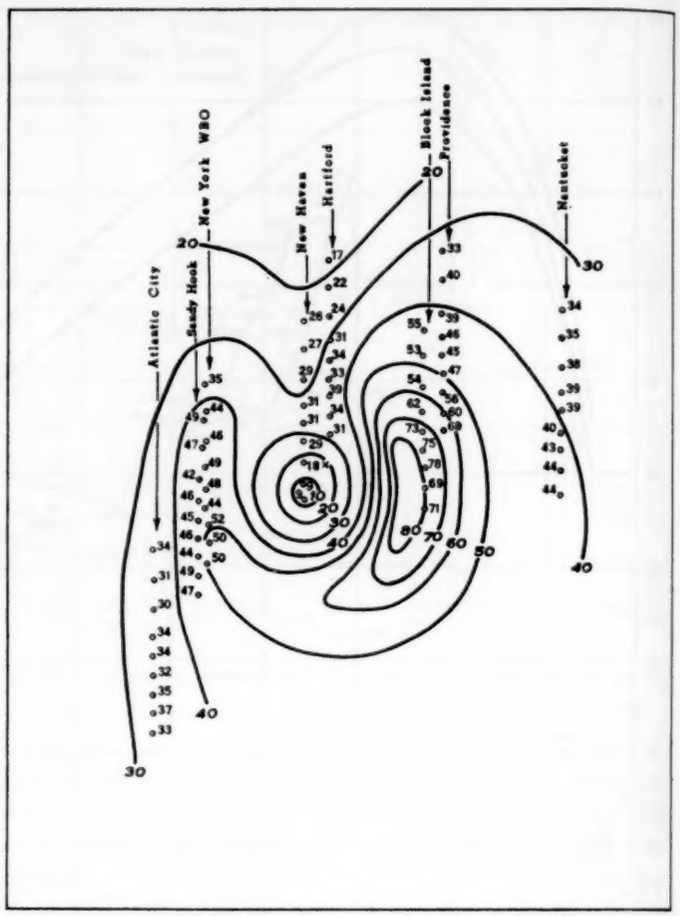

FIGURE 9.—Surface wind speeds (adjusted to 30 ft. off-water and 1500 EST intensity of storm) plotted relative to pressure center X, 1400–1600 EST, September 21, 1938. Speeds are in miles per hour. Distance scale same as figure 10.

cerned and 1500 EST, the reference time of the composite pattern. This was done for each time.

Ship and coastal wind speeds at the time of each map were plotted and the adjusted composite patterns were further modified to fit these. The coastal data were adjusted only for the reduction to the standard 30-ft. elevation. The 1200 EST isotach map is similar to Hughes' [5] mean pattern even though Hughes' data were south of 30° N. and did not include storms that had recurved.

6. WIND DIRECTION

Deflection angles (the angle between the wind direction and a tangent to a circle about the wind center) were examined in some detail in an effort to find a pattern through the storm as a whole that could be extrapolated from the regions of data to the regions of no data. Since the pattern was not well defined, the deflection angles for another great New England storm, that of September 1944, were added to expand the data. Average angles by zones from both New England storms were plotted on the same chart, together with the mean of the two storms (fig. 11), and a rough analysis was drawn to the data. This pattern is only approximately indicative of what would be observed in a hurricane over the ocean because of the effects of local topography, variations in friction from sea to land and from one place over land to another, height of the wind vane, inadequacy of 8-point wind-direction measuring systems, etc., in addition to the inherent variability of the wind direction. A large scatter in deflection angles seems to be typical for hurricanes. Johnson [9], working with high quality wind-direction data, showed a large scatter of the deflection angles throughout the August 1949 hurricane out to 70 miles from the center.

It appears from the right half of figure 11, and from Hughes' [5] mean wind-direction pattern, that use of a mean deflection angle of 25° outside R and 20° inside R would be satisfactory in correlating winds with surges in the 1938 storm. It also follows that, lacking a model of wind directions with a more substantial empirical or theoretical basis, further refinement of the wind direction along the critical fetch of maximum wind speeds in the shoreward quadrant of the storm is not warranted.

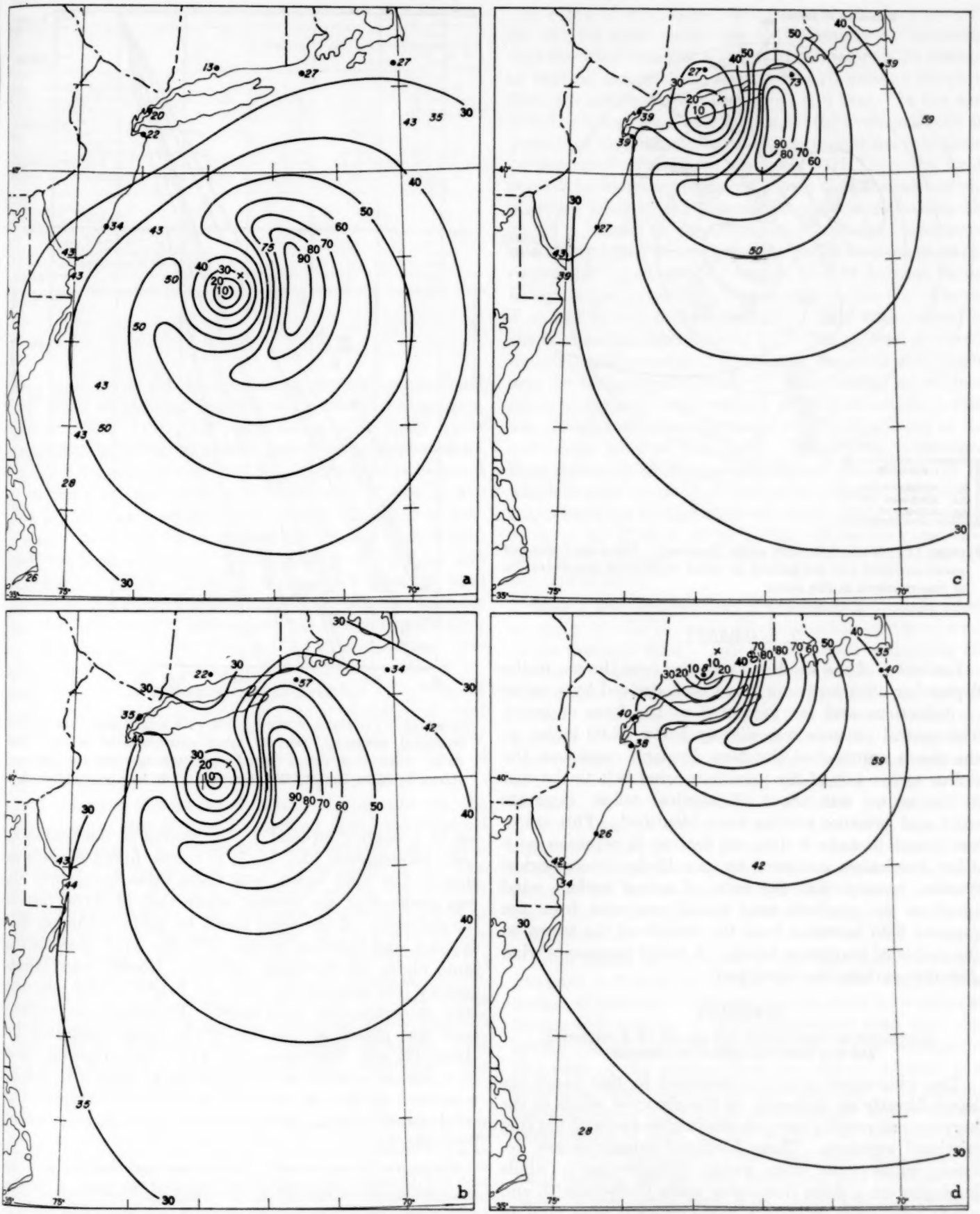

FIGURE 10.—30-foot wind speeds (m. p. h.) September 21, 1938. (a) 1200 EST, (b) 1400 EST, (c) 1500 EST, (d) 1600 EST. Data with dot positions are Weather Bureau Station observations, reduced to 30 feet. Data without dot positions are ship reports, unadjusted. X shows location of pressure center.

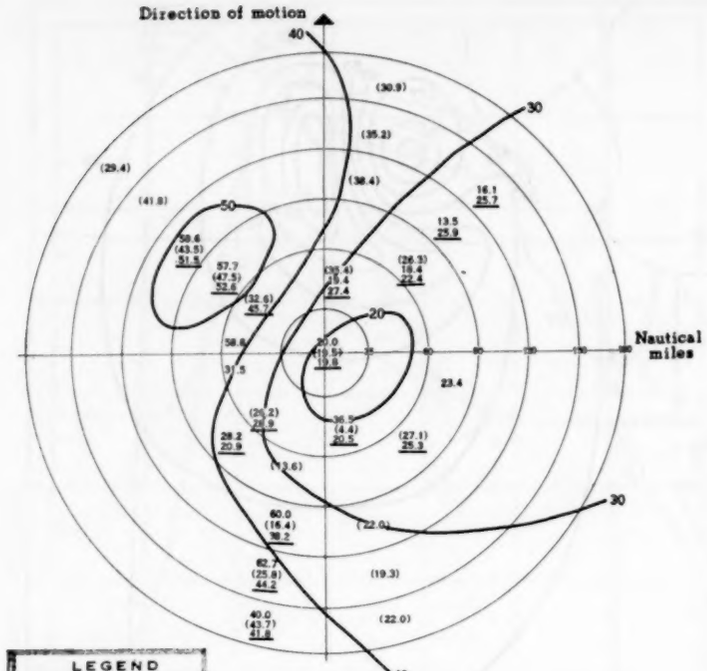

FIGURE 11.—Wind deflection angle (degrees). Data are means of zones outlined and are plotted at point of greatest concentration of observations in the zones.

7. SUMMARY

Estimates of the wind and pressure over the sea in the September 1938 hurricane have been derived by a series of deductions and are presented in the form of maps. The central pressure was slightly below 28.00 inches as the storm approached the New England coast and the radius to the axis of the maximum wind belt to the east of the center was about 50 nautical miles. Separate wind and pressure centers were identified. This storm was found to have a dynamic feature in common with other hurricanes analyzed by the Hydrometeorological Section, namely that the ratio of actual surface wind speed to the gradient wind speed computed from the pressure field increases from the outside of the storm to the radius of maximum winds. A rough composite wind direction pattern was developed.

APPENDIX

ADJUSTMENT OF WIND SPEEDS FOR HEIGHT OF ANEMOMETER AND FOR DIFFERING FRICTIONAL SURFACES

The wind-speed analyses presented in this paper are based directly or indirectly on the observed winds in the hurricane at varying anemometer heights and with varying frictional exposure. Three frictional categories are defined, winds over open water ("over-water"), winds impinging on a shore from open water ("off-water"), and winds over land, including winds blowing from land to sea at a coast ("off-land"). In studies of hurricanes at Lake Okeechobee, Fla., [6, 7, 8] it was found feasible to stratify the wind speeds into these three categories. It was assumed in the present study that at 1,000 meters the reduction of the wind speed by surface friction disappears, and therefore at this level the wind speed is the same above all frictional surfaces. Rossby and Montgomery [15] have shown that the height of the gradient level increases with wind speed. Therefore the gradient level was placed at the top of the range suggested by Byers [2] and Petterssen [10, 11]. Experimental data from various authors on the variation of wind with height were then applied to construct curves of the variation of wind speed from the surface to the 1,000-meter gradient level (fig. 12).

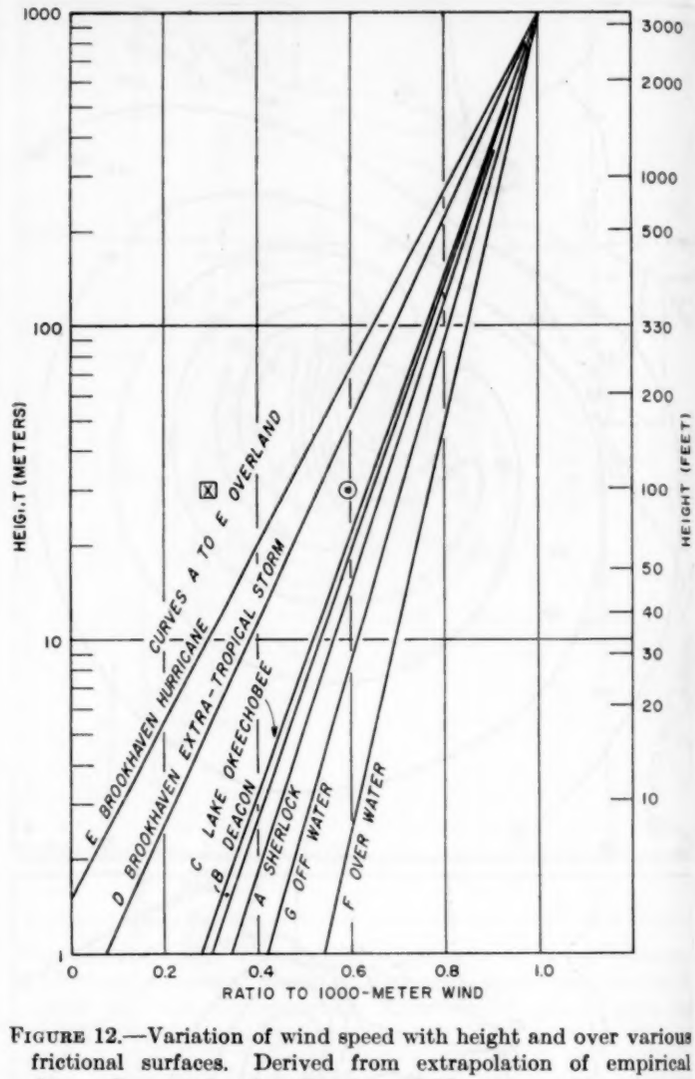

FIGURE 12.—Variation of wind speed with height and over various frictional surfaces. Derived from extrapolation of empirical data. Computed ratios (Rossby and Montgomery formula) are shown by ⊠ for rough, hilly country and ⊙ for open grassland.

Deacon's [4] summary of experimental determinations by various investigators of the ratio of winds at a height of 100 meters to the 10-meter wind is reproduced in table 1.

TABLE 1.—*Ratios of 10-, 100-, and 1,000-meter wind speeds*

	V_{100}/V_{10}	V_{1000}/V_{10}	V_{10}/V_{1000}	Curve in fig. 12
Sale, Victoria, Australia	1.44	1.88	0.53	B
Cardington, Bedfordshire, England	1.48	1.96	.51	
Leafield, Oxfordshire, England	1.48	1.96	.51	
Quickhorn, Germany	1.71	2.42	.41	
Brookhaven Laboratory, Long Island, U. S. A. (extratropical storm).	1.83	2.66	.38	D
Brookhaven Laboratory (hurricanes)	2.20	3.40	.29	E
Akron, Ohio, U. S. A.	1.66	2.32	.43	
United States (Sherlock's data)	1.39	1.78	.56	A

Source of ratios:
V_{100}/V_{10}: observed.
V_{1000}/V_{10}: extrapolated from V_{100}/V_{10} by use of logarithmic law:
$V_{1000}-V_{100}=V_{100}-V_{10}$
$V_{1000}/V_{10}=2\,(V_{100}/V_{10})-1$
V_{10}/V_{1000}: reciprocal of V_{1000}/V_{10}

For sources of observations, see Deacon [4], except for Brookhaven (hurricanes) which is from [17].

Also included is a ratio determined for hurricanes Carol and Edna of 1954 at Brookhaven National Laboratory, Upton, N. Y. [17]. All observations were made under conditions of either adiabatic lapse rate or moderate-to-high wind speeds. Smith and Singer [17] recently reported that the wind speeds at four levels from 37 feet to 410 feet on the tower at the Brookhaven Laboratory in hurricanes Carol and Edna showed a variation with height that fits both a logarithmic law ($V_1-V_2=K \log z_1/z_2$) and a power law ($V_1/V_2=(z_1/z_2)^n$), where the V's are the wind speeds at the corresponding heights, z. The logarithmic law was employed to extrapolate the experimental values of variation of wind with height up to the assumed gradient level of 1,000 meters and to compute ratios of 10-meter winds to 1,000-meter winds (table 1). Selected ratios were used to construct curves of variation of wind speed with height (fig. 12, curves A, B, D, and E). These are all curves for winds over land. The development of an additional off-land curve, C, is described later.

Of the wind factors listed in table 1, only one set, the second set of Brookhaven Laboratory data, was obtained in hurricanes. The others were measured at lower wind speeds, in the range from 20 to 45 or 50 m. p. h. The question arises of the applicability of the lower speed results to hurricanes. In hurricanes Carol and Edna at the Brookhaven National Laboratory there was no significant change in the ratio of the 37-ft. wind speed to the 150-ft. wind speed over the range of observed speeds, from 31 m. p. h. to 79 m. p. h. at the 150-ft. level. This evidence was given the greatest weight and it was assumed that the determinations of wind-speed variation with height in the lower speed range are reasonably applicable to hurricanes at similar sites.[5]

[5] There are other, and conflicting, indications on whether the ratio of the wind speeds at two heights varies with the wind speed. Comparison of curves D and E of figure 12 shows that at the Brookhaven site there was a greater relative variation of wind speed with height in the two hurricanes (curve E) than in the lower-speed extratropical storm (curve D). On the other hand, figure 30 of *Hydrometeorological Report* No. 32 [8] shows that the ratio of off-land wind speed to over-water speed at Lake Okeechobee increased as the wind speed increased. A logical deduction from this observation is that the relative variation of wind speed with height is less for higher speeds.

A curve of wind-speed variation with height (curve F, fig. 12) for open water was drawn simply by assuming that the wind speed at a height of 10 meters is 70 percent of that at the gradient level. The 70 percent factor is from the statement by Petterssen [10] that "At the surface, the actual wind over land is, on the average, about 40 percent of the geostrophic wind, whereas at sea it is about 70 percent." Petterssen also said [11] ". . . the analysis of the weather charts has shown that the ratio of the observed wind to the geostrophic wind is approximately as 2:3." Curve G, for off-water speeds, was constructed so that the speed at a height of 40 feet is 89 percent of the over-water speed at that height, the 89 percent factor being obtained from Lake Okeechobee studies [8]. Curves F and G appear to be reasonably placed with respect to the land surface curves (A to E). This indirect approach to constructing curves of wind-speed variation with height may be tested qualitatively by constructing an off-land curve in the same way. Thus, curve C for off-land wind was drawn such that the 40-ft. wind is 76 percent of the over-water wind at that level. That factor is obtained from figure 30 of *Hydrometeorological Report* No. 32 [8] which depicts ratios of off-land to over-water wind and is taken from the highest observed wind speeds (over-water, 81 m. p. h.; off-land, 62 m. p. h.). It can be noted that this curve lies very close to the curve for Deacon's Sale site (curve B).

Rossby and Montgomery [15] develop a formula (their equation 35) for the ratio of the anemometer-level wind to the gradient wind. This was not used in deriving the curves of figure 12 because knowledge is required of factors unavailable, such as the roughness parameter, z_0, and the angular difference in wind direction at anemometer and gradient levels. However some interesting comparisons can be made. Those authors obtained two typical solutions to the formula by substituting reasonable values: a ratio of 0.595 for open grassland and 0.293 for rough hilly country, both at a wind speed of 15.6 m. p. s., and an anemometer height of 30 meters. These ratios are plotted on figure 12 and compare well with the other data. The point for the hypothetical "rough hilly country" correctly lies to the left of curve E as the Brookhaven site is less rough than this.

From figure 12 the ratio of the wind speed over any frictional surface at any height to that over any frictional surface at another height may be obtained by reading off the respective ratios to the gradient-level wind and then taking the quotient of these ratios. (The variation in character of the surface from one site to another is so great that these ratios are far from precise. The principal justification for their use is that they are better than no adjustments at all). In the analysis of the 1938 hurricane, Block Island and Nantucket were considered off-water stations for all wind directions, and Atlantic City, Sandy Hook, New York WBO, and New Haven either off-water or off-land, depending on wind direction. Provi-

dence was off-land except over a very restricted sector in the direction of Narragansett Bay. Curve B of figure 11 was used for all off-land adjustments (arbitrary choice).

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THE WEATHER AND CIRCULATION OF JULY 1956[1]

Including Some Aspects of Momentum Flux in Relation to an Intense Polar Vortex

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1. WEATHER AND CIRCULATION IN THE UNITED STATES

The prevailing westerlies of middle latitudes were north of normal over eastern North America during June 1956 [1]. During July, however, they shifted southward again, to a latitude more typical of May (compare fig. 6 of [1] with fig. 1A of this article). This abnormal southward displacement of the westerlies brought below normal temperatures and frequent cyclonic activity, accompanied by prolonged cloudiness and precipitation, to the northeastern quarter of the nation.

This July was the coldest in over half a century for eastern Pennsylvania, and parts of New York State, New England, and Michigan. In the Northern Plains monthly mean temperature anomalies of 4° F. below normal were the lowest in over forty years at several stations. At some localities in the northern part of the country afternoon temperatures barely climbed above 80° F., while at Sault Ste. Marie, Mich., the highest temperature was only 78° F.—the lowest for any July. Minimum temperatures in the low forties were not uncommon, with some rural areas in the lower peninsula of Michigan even reporting frosts.

These subnormal temperatures were for the most part situated in an area of below normal 700-mb. heights (fig. 2), just north of the mean frontal position (fig. 3) and the largest wind speed anomalies (fig. 1B). Compared with July 1955 [2], this month was diametrically opposite in circulation pattern over the eastern part of the country. Then, the strongest westerlies, mean frontal zone, and principal storm track were displaced north of the Canadian border, while temperatures ranged as high as 4°–6° F. above normal.

July 1956 was also one of the cloudiest and rainiest on record for much of the northeastern quarter of the nation with many stations reporting rain (including trace) on 20 or more days during the month. Note the large number of days with fronts (as many as 23) over Kentucky and Virginia in figure 3. Yet, monthly rainfall totals for some communities were not unusually great and, in a few places, were even subnormal. For example, at Scranton, Pa., there occurred 23 days with rain, but the monthly total of 3.18 inches was 2.15 inches below normal.

The frequent storminess responsible for this cool, rainy weather can be seen from the cyclone tracks of Chart X.

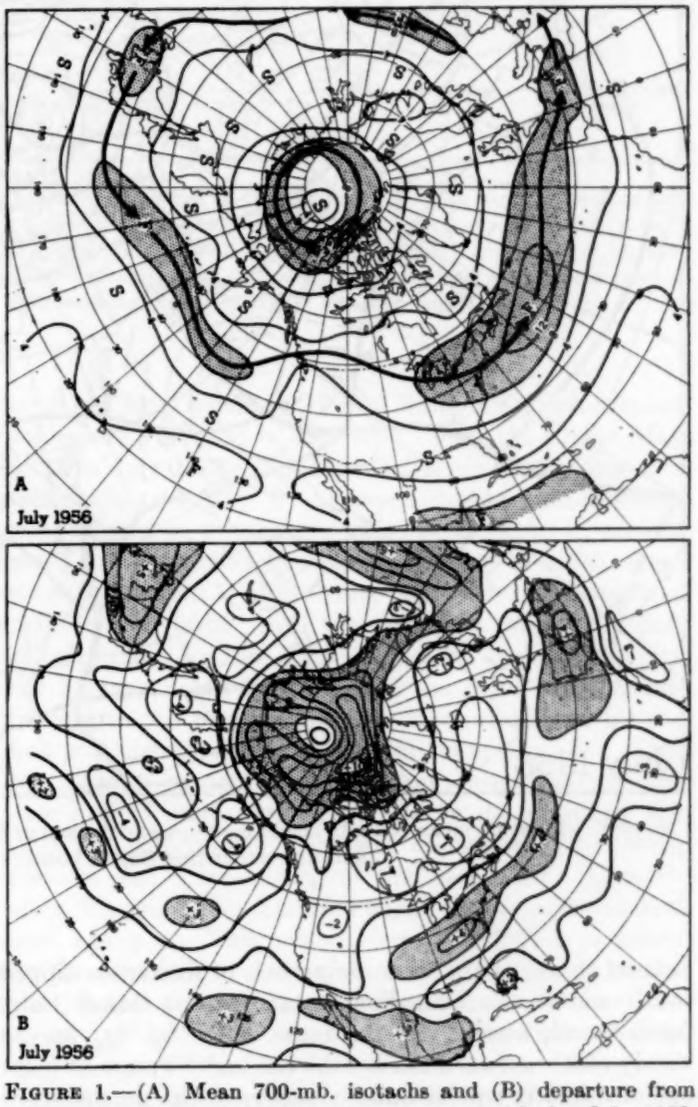

FIGURE 1.—(A) Mean 700-mb. isotachs and (B) departure from normal wind speed (both in meters per second) for July 1956. Solid arrows in (A) indicate axis of the mean jet stream at 700 mb. Wind speeds in (A) greater than 8 m. p. s., and wind speed anomalies in (B) greater than 4 m. p. s. are shaded. Note the strength of the circulation in the Arctic.

[1] See Charts I-XVII following p. 289 for analyzed climatological data for the month.

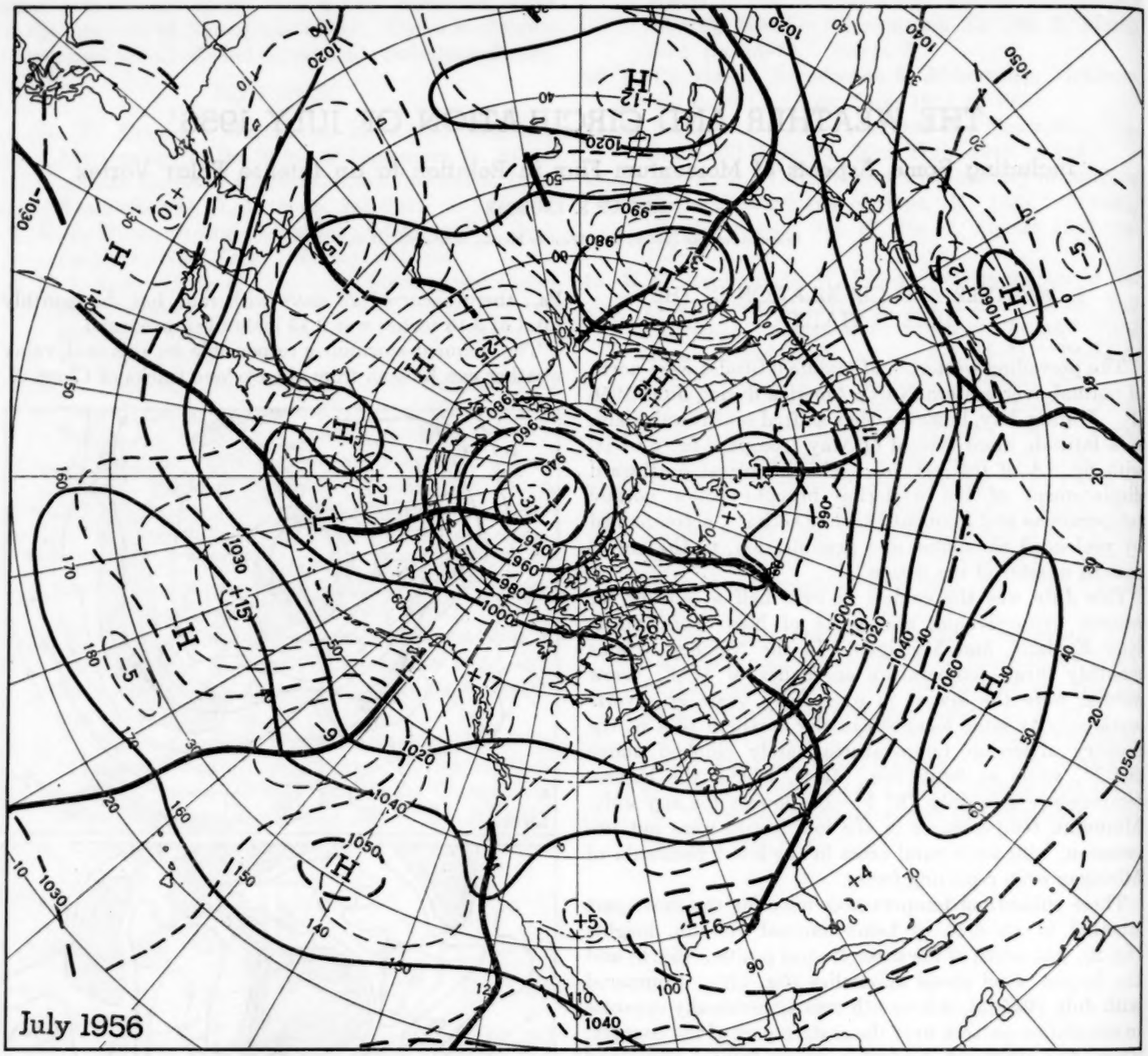

FIGURE 2.—Monthly mean 700-mb. height contours and departure from normal (both in tens of feet) for July 1956. The deep polar vortex and its surrounding ring of positive height anomalies are notable features.

Several times, as the meandering belt of westerlies dipped south across the Canadian border, deep, closed Lows formed and, steered by the mean flow (fig. 2), moved slowly southeastward toward the Lakes.[2] Frequent passage of migratory polar anticyclones over the northeastern quarter of the country (Chart IX) was also responsible for cool weather in that sector.

In contrast to the Northeast, warm, dry weather accompanied above normal 700-mb. heights over Texas and Louisiana as well as in the extreme western part of the country (Chart I-B and fig. 2). Fort Worth, Tex., with an average maximum temperature of 101° F. and a monthly mean of 89° F., reported this month as the second hottest on record.

Below normal temperatures over the Southwest can in part be related to the rather general cloudiness and rainfall that occurred (Charts III, VI, VII). However, some of the cooler days received the maximum possible sunshine. This was generally true early in the month when the moisture content of the air was low and a strong diurnal range of temperature occurred with low nighttime minima. Stronger than normal northwesterly flow on

[2] See article by McQueen and Martin in this issue for a discussion of one such synoptic situation.

the monthly mean sea level map (Chart XI and inset) was also indicative of cool weather in this area.

2. CIRCULATION AND WEATHER IN OTHER PARTS OF THE NORTHERN HEMISPHERE

The mean circulation for July 1956 was characterized by stronger than normal westerlies in the Arctic and in mid-latitudes with subnormal values in the zone between (fig. 1). This latitudinal distribution of wind speed was associated with a belt of above normal 700-mb. heights completely circumscribing an unusually intense polar vortex (fig. 2). Positive height anomalies in this ring included centers of 270 ft. over northeastern Siberia and 260 ft. over northern Canada. These were indicative of the persistent blocking anticyclones that formed within this belt. The polar vortex, with an extreme height anomaly of −310 ft., was a site of frequent and sometimes intense cyclonic activity. On its periphery, mean 700-mb. wind speeds reached values as high as 12 m. p. s., or 10 m. p. s. above normal (fig. 1).

Cyclonic activity was also typical of the regions of subnormal 700-mb. heights south of the circumpolar ring of high pressure, and Lows forming in these regions reached unusual intensities for this time of year and these latitudes. This month, with its large, deep, and often slow-moving vortices over the east-central Pacific, the Great Lakes region, the British Isles, and Russia, seemed more typical of spring than summer.

The largest negative height anomaly (−310 ft.) was that over Russia, where a large closed Low center was present even on the monthly mean 700-mb. map (fig. 2). Some of the coldest temperatures with respect to normal for the lowest 300 mb. of the troposphere occurred here as well as considerable rainfall. A trough extending from this center southward to the Persian Gulf was particularly well marked over Iran, where it caused flood-producing rains.

Heavy rains also fell in a center of negative height anomaly (fig. 2) over the British Isles. This precipitation was associated with numerous storms travelling across the Atlantic on a track depressed south of normal (Chart X). These rains were reported as being the heaviest in 80 years—an unusual contrast to May which was the driest since 1860.

In the eastern Pacific a trough and below normal heights appeared in a region that is climatologically favorable for ridges. Only one other July (1950) in the past 22 years has had a trough in this location. Development of this trough accompanied rising 700-mb. heights over the Bering Sea and Alaska as blocking shifted westward from Canada. Concurrently there was a slow weakening of the climatologically favored trough along the west coast of North America. This latter trough did not reintensify until the Pacific High was again established at the end of the month.

3. EVOLUTION OF THE JULY CIRCULATION

The development of this month's circulation is well portrayed by a series of 15-day mean zonal wind speed profiles (0° westward to 180° long.) from the period May 16–30 to July 16–30 (fig. 4). Reference to the first two graphs (fig. 4A and B) shows a jet at mid-latitudes strengthening to an average value of 12 m. p. s., accompanied by a marked increase in the north-south shear across the current. This strong zonal flow, however, was unstable and "split" into two branches during the first half of June. At this time a locus of cyclonic activity near Novaya Zemlya shifted into the polar cap—its impression on the mean zonal flow being indicated by the weak polar jet appearing at high latitudes (fig. 4B). This polar vortex underwent significant intensification the latter half of June (fig. 4C), while on its periphery the increasing anticyclonic shear was accompanied by rising pressures and a weakening of the zonal flow. Concurrently, the jet at mid-latitudes edged south and weakened about 3 m. p. s. This pattern, once established, persisted into the first half of July with a slight weakening of the polar jet and a strengthening and southward shift of the one at mid-latitudes (fig. 4D). By the end of July (fig. 4E) a significant weakening of both jets took place, accompanied by increasing westerlies near 60° N. as the two currents approached one another. Over the Middle Atlantic States, however, the westerlies continued strong with no improvement in the weather.

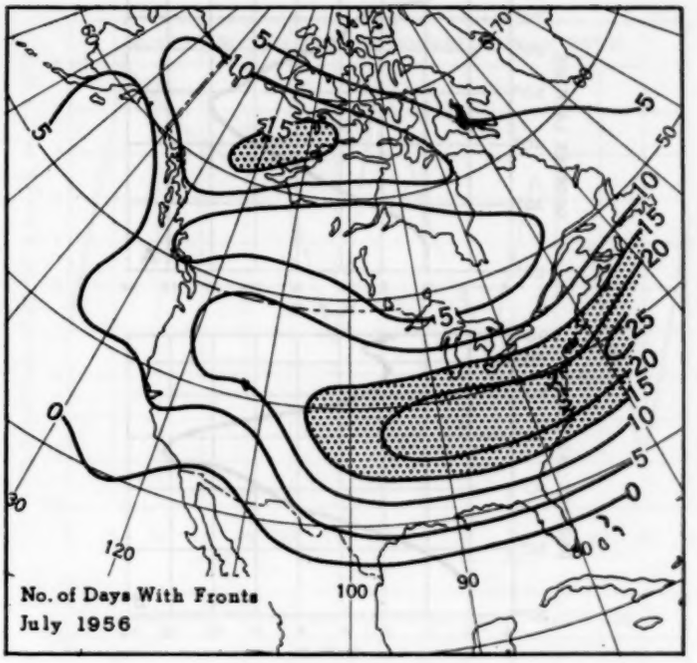

FIGURE 3.—Number of days in July 1956 with surface fronts of any type (within squares with sides of approximately 500 miles). Frontal positions taken from *Daily Weather Map*, 1:30 p. m. EST. The unusually high frequency of fronts over the eastern part of the country was responsible for cool, rainy weather.

4. MOMENTUM TRANSFER AT HIGH LATITUDES

The maintenance of westerlies in any zonal ring requires a transfer of angular momentum into that ring to compen-

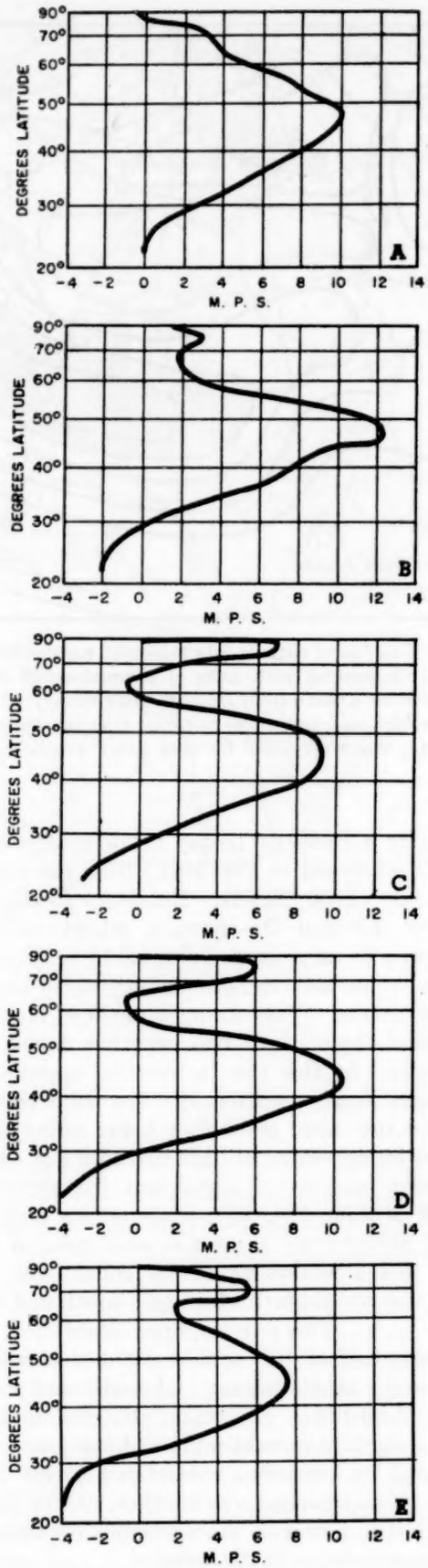

FIGURE 4.—15-day mean zonal wind speed profiles for the Western Hemisphere at 700 mb. for (A) May 16–30, (B) May 30–June 13, (C) June 16–30, (D) June 30–July 14, and (E) July 16–30, 1956. Note development of double jet in June and its persistence through July.

sate losses due to friction. This momentum is supplied by the surface easterlies (essentially the trade wind belt) which receive westerly momentum from frictional interaction with the earth [3]. Variations in the zonal momentum at a given level result from the effects of: (1) horizontal convergence of relative angular momentum, (2) vertical convergence, (3) mean meridional circulations, (4) friction, and (5) mountain torques. Of these, the first has been considered one of the most important sources for the westerlies. It is also the most easily measured and can be computed from constant pressure charts using the geostrophic assumption.

Considering the strong westerlies in the Arctic and in mid-latitudes this month with blocking activity in the zone between, it was thought interesting to compare fluctuations of the zonal momentum for various belts with the horizontal flux of relative angular momentum into these belts. This was carried out for the polar cap (region north of 75° N.) and the zonal ring from 55°–75° N. Time did not permit a more extensive study.

Starting with the relationship [4]

$$T = \int_0^{2\pi} \rho R^2 \cos^2 \varphi \, uv \, d\lambda \tag{1}$$

where T_φ is the horizontal transfer of relative angular momentum across latitude φ, R is the earth's radius, ρ is the density of the air, u and v are respectively the eastward and northward components of the horizontal wind flow, and λ is the longitude, one can employ the geostrophic assumption and evaluate the above integral numerically to obtain

$$T_\varphi = K \sum_{\lambda=0}^{360} (\Delta z)_\varphi (\Delta z)_\lambda \tag{2}$$

where K is a constant for the latitude in question, and $(\Delta z)_\varphi$ and $(\Delta z)_\lambda$ are the north-south and east-west contour height differences respectively. For this study these were obtained from heights 5° north and south (or east and west) of the given point, and the summation of products was then taken around a parallel of latitude. The data used were 5-day mean 700-mb. heights centered four or five days apart during June and July. Such computations have been made on 5-day mean 700-mb. charts previously for other cases [5, 6]

The results of applying equation 2 at 75° N. are shown in figure 5 (top). For comparison, the zonal wind speed for the polar cap, where a mean Low was located from June 4 to the end of July, has also been plotted in figure 5 (bottom). This figure indicates that: (1) During this period the convergence of momentum into the polar cap was essentially positive, averaging about 4×10^{18} gm. cm. sec.$^{-2}$; (2) The long-period trend of zonal wind speed in the polar cap, and that of momentum flux across 75° N. were both downward, especially during July; and (3) Variations in the two curves can be related. This would suggest that the polar vortex was largely maintained by a flux of momentum across its boundaries, which was

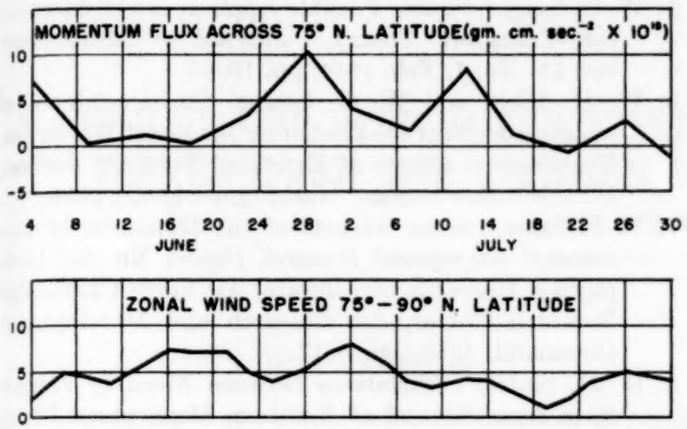

FIGURE 5.—Upper curve.—Momentum flux at 700-mb. level across 75° N. (gm. cm. sec.$^{-2}\times10^{18}$). Lower curve.—Zonal wind speed at 700 mb. for the belt 75°–90° N. (m. p. s.). Momentum flux into the polar cap was essentially positive.

probably achieved by the numerous travelling cyclones that moved into the Arctic during this period. Tracks of several of these are portrayed in Chart X, and many others were also noted in the Eastern Hemisphere.

Agreement between the two curves is poorest about June 16 and July 22. At these times there occurred strong responses of the zonal wind speed to small values of the momentum flux. This points to other important sources of momentum for the 700-mb. level, such as vertical momentum convergence, effects of mean meridional circulations, or mountain torques due to the Greenland plateau.

The best agreement between the two curves is between June 24 and July 15, which may be a period when the contribution from other sources was least. During this period both maximum and minimum values of zonal wind speed $\left(\frac{\partial u}{\partial t}=0\right)$ occurred with a value of the momentum flux of approximately 5×10^{18} gm. cm. sec.$^{-2}$. This may perhaps represent a rough measure of friction for this interval.

A better method of studying the polar vortex would be to calculate the momentum flux across the boundaries of the vortex itself, as suggested by Starr [7], rather than across fixed latitudes. However, time did not permit this to be done here.

The next zone that was examined was the 55°–75° N. latitude belt which, it will be recalled, was one of high pressure completely surrounding the pole on the July mean map (fig. 2). As a measure of the momentum convergence into this belt, the difference between the flux across 55° N. and that across 75° N. was obtained (fig. 6, top). Comparing this curve with the zonal wind speed for the belt (fig. 6, bottom) reveals fair agreement, with periods of relatively strong momentum convergence into the belt accompanied and followed by stronger zonal wind speeds, and periods of divergence (negative values of the flux difference) by weak winds. The periods of

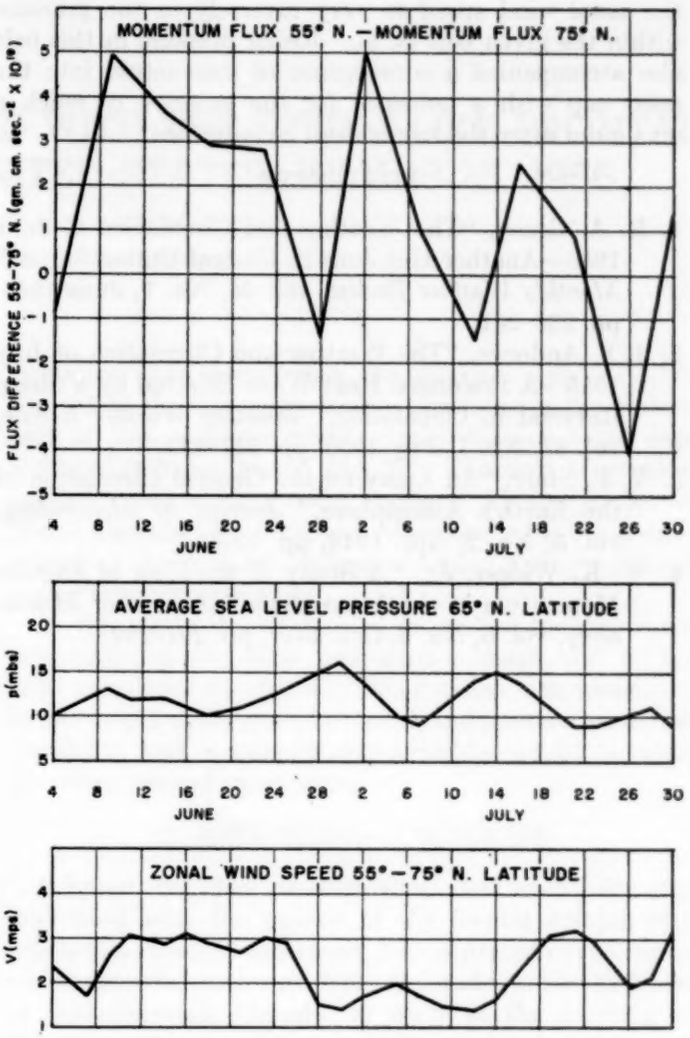

FIGURE 6.—Upper curve.—700-mb. momentum flux at 55° N. minus that at 75° N. (gm. cm. sec.$^{-2}\times10^{18}$). Middle curve.—Average sea level pressure (mb.) at 65° N. Lower curve.—700-mb. zonal wind speed (m. p. s.) 55°–75° N. Note tendency for small and negative values of the convergence to be accompanied by rising pressure and decreasing wind speeds.

strong divergence of momentum in the 55°–75° N. belt (note sharp dips in fig. 6, top, on June 28, July 12, and July 26) were periods of strong momentum accumulation in the polar cap (note sharp peaks in fig. 5, top). However, only about 40 percent of the momentum lost by the 55°–75° latitude belt was transported north across 75° N. The remainder was transported southward across 55° N. into the belt of strongest westerlies. Of interest is the long-period downward trend of momentum convergence during July and the corresponding upward trend in the zonal motion for the belt. This again points to other important sources of momentum.

The variation in sea level pressure within the 55°–75° N. belt (fig. 6, middle) shows momentum flux divergence accompanied by increasing pressure, and convergence accompanied by decreasing pressure. This relationship is expected, since it is well known that a tendency exists for

the zonal wind speed to vary inversely as the pressure within the given belt [8, 9]. Rising pressure in this belt also accompanied a convergence of momentum into the polar cap with a tendency for the pressure to reach a maximum after the momentum convergence.

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A COOL, DAMP PERIOD ASSOCIATED WITH THE SPRING-LIKE FRONTAL WAVES OVER THE EASTERN UNITED STATES, JULY 18-25, 1956

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1. INTRODUCTION

The occurrence of well-defined polar fronts in the Gulf Coast States during the month of July, while not necessarily a rarity, has been at least an unusual happening during the past several Julys. Upon reference to the Historical Weather Maps [1] it was noted that there have been Julys when as many as five polar fronts have swung southward over these Gulf States, while, on the other hand, there have been other Julys when the southern limit of the polar fronts has remained north of the 35th parallel. However, in most of the cases observed there was an ephemeral duration of the front in these southern latitudes; certainly few had the longevity of the case of July 18–25, 1956, the subject of the present study. That such prolonged activity should occur during a summertime frontal condition may in part have been due to the fact that temperatures in the airmass to the north of the front were considerably lower than the mean temperatures as shown by Showalter [2] in his study of summer airmass properties. And also that several stations in the southern portion of the country recorded temperatues in the upper air that were equal to or below the minimum values presented in the paper on "Extreme Temperatures in the Upper Air" [3].

Surface temperatures over the Eastern States during these situations can and often do vary greatly from year to year. This dependency it appears is closely related to the pressure distribution to the north of the front rather than to the frequency of frontal occurrence, or in other words to the resulting flow of air and the source region of the cool air. In the present study the persistency of the cold air (see fig. 1) was due to the formation and stagnation of a cold Low over the western Great Lakes region during the early portion of this period. Thus, while located in the Great Lakes region the Low could continue to reinforce the already cool air over the East by a constant flow of Canadian air from the Hudson Bay region, one of the principal source regions of cool air during the summer months in North America.

Precipitation associated with these fronts is frequently heavy in portions of the Gulf Coast and the Atlantic Coast States, and in this regard the July 1956 situation was no exception. However, the frequency of precipitation was even more outstanding with over 80 percent of the first-order weather stations in the eastern half of the country recording rain four or more days during this 8-day period (fig. 2). The area of greatest rainfall during this period occurred in the States of North Carolina, Virginia, and Maryland with the greater portion of the total having been obtained during a 12-hour period between July 19–20.

In this study it was decided to confine the discussion principally to the 0030 GMT surface maps and to the 0300 GMT upper air charts. This decision was occasioned by the length of the period covered, and because maximum instability and maximum temperatures would be most prevalent nearest those times.

2. ANTECEDENT CONDITIONS

A broad, flat, low pressure area, that eventually was associated with the genesis of the frontal system with which this article is concerned, first appeared in the region which extends northward from Colorado to the Province of Saskatchewan, Canada, as early as the morning of July 11, 1956. In the next 24 hours this broad trough moved eastward as it extended from a point just south of Lake Winnipeg, Canada into northern Mexico. Within the next 24 hours a 1007-mb. Low developed at the surface over central Wisconsin attended by frontogenesis. In the upper air, prior to the 13th, the 500-mb. chart presented a moderately deep Low centered over the eastern shore of Hudson Bay with a rather sharp trough that extended westward from James Bay to northern Lake Winnipeg. The maximum cyclonic vorticity associated with this upper trough was advected to the vicinity of the surface Low by the morning of the 13th to aid in the development and deepening of this new cyclonic center.

During the next 48 hours the central surface pressure lowered some 12 mb. as the Low moved northeastward through the St. Lawrence River Valley to the extreme eastern portion of Quebec Province. In this latter position it produced a northerly flow of air extending from the center of the Low to the axis of the High which was located along a line from the extreme western Great Lakes to the western portion of Hudson Bay. This northerly flow resulted in cold air pouring southward across most of

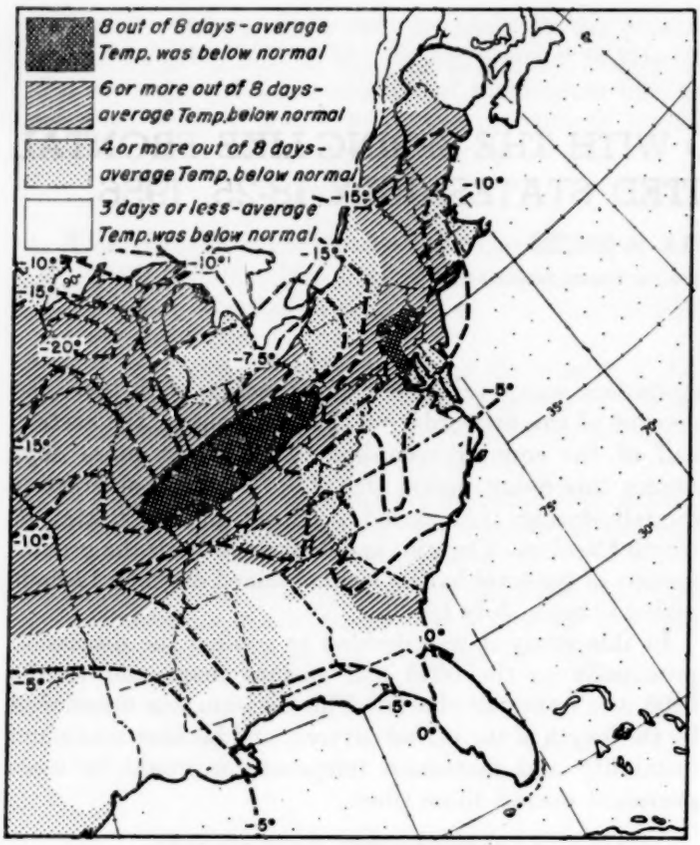
8 out of 8 days - average Temp. was below normal
6 or more out of 8 days - average Temp. below normal
4 or more out of 8 days - average Temp. below normal
3 days or less - average Temp. was below normal

FIGURE 1.—Number of days during the 8-day period July 18–25, 1956 when daily average temperature was below the normal July value; total days of occurrence are indicated by different types of shading. Isolines show the largest one-day maximum temperature negative departure from normal July maximum. (Based on data from first-order stations.)

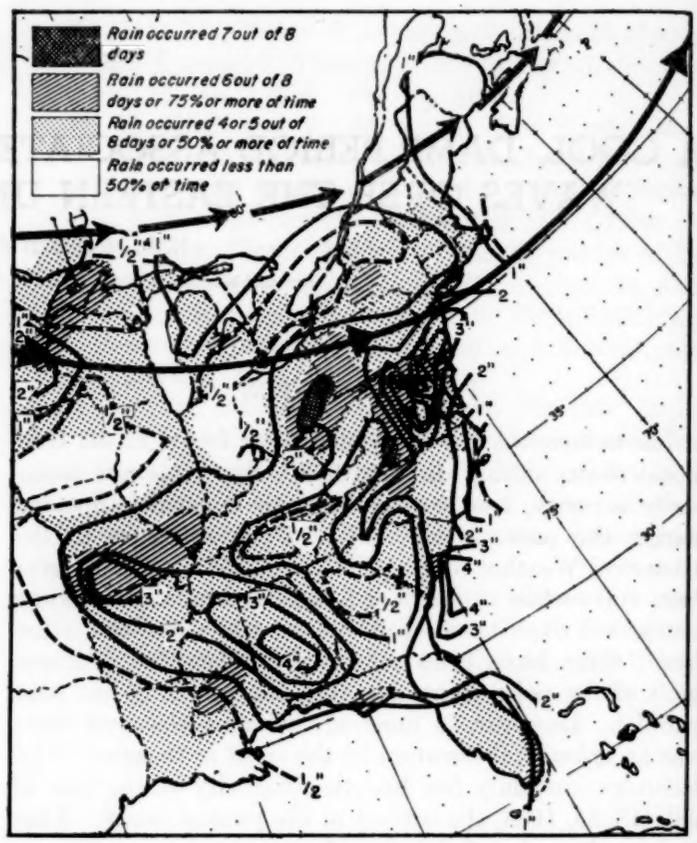
Rain occurred 7 out of 8 days
Rain occurred 6 out of 8 days or 75% or more of time
Rain occurred 4 or 5 out of 8 days or 50% or more of time
Rain occurred less than 50% of time

FIGURE 2.—Number of days during the 8-day period July 18–25, 1956 when a trace or more of precipitation occurred during a calendar day. Shading of different types has been used to distinguish different totals. Also shown are isohyets (continuous lines at one-inch intervals) for the period; isohyets in excess of 6 inches are omitted due to size of area. (Based on data from first-order stations.) Continuous, heavy, arrowed line shows average position of July 1956 700-mb. jet stream in comparison to normal July 700-mb. jet stream (broken, arrowed line).

northeastern United States and the intensification of the front as it stretched southwestward across the central Atlantic Coast States, the Ohio Valley, central Mississippi Valley, and into the Northern Plains.

In eastern Montana cyclogenesis was beginning to occur in conjunction with a 500-mb. trough over northern Alberta. Northwest of this new Low, pressures were rising at all levels in association with a building upper-level ridge.

By the morning of July 17, the surface trough extended southwesterly along the Atlantic Coast to the Virginia Capes, then westward to extreme southern Illinois where branching of this trough occurred; one arm extended northwestward to Lake Winnipeg, the other arm west-southwestward to central Oklahoma thence southwestward across central Texas. The surface front was in the main trough and followed the southern arm from the point of branching.

Associated with this surface picture was a broad upper-level trough over most of the area east of the Mississippi River. A major upper-level ridge was located along the west coast of the United States and Canada while a secondary ridge prevailed from western Texas to North Dakota. A small closed 500-mb. Low that had migrated from the northern portion of the Gulf of Alaska eastward along the 60th parallel started to curve sharply southward on the morning of July 17 and was centered near Reindeer Lake, Saskatchewan. The maximum cyclonic vorticity associated with this Low was to the north of Edmonton, Alta., in a favorable position to be advected over the surface trough near Lake Winnipeg, Sask. This presented a situation similar to that of the previously mentioned Low which had intensified rapidly over the Great Lakes region, except that in this present circumstance a strong band of westerlies prevailed across the west coast of the United States where previously a trough had existed.

3. SYNOPTIC SITUATION

In the early morning of July 18 (GMT) a cold front of moderate intensity was east of Sable Island, N. S., and extended in a southwesterly direction to near Wilmington, N. C., thence westward across the northern portion of the

Gulf Coast States, and terminated over north-central Texas (fig. 3A). Over the interior of the Nation the category of the front had been classified as weak on the basis of the gradient of the 1000–500-mb. thickness lines to the rear of the front. Because of widespread thunderstorm activity over the Southeastern States, the surface temperature discontinuity across the front was poorly defined in that area, but west of the Mississippi River surface temperature differences of from 12° to 20° F. were present.

The north-south orientation of the High near Lake Superior in conjunction with the Low over Labrador, Canada, insured a continued flow of cool Canadian air from the Hudson Bay region spreading over the Eastern States north of the surface front.

The 500-mb. constant pressure chart for 0300 GMT, July 18 (fig. 3B) presented a well-defined trough along the eastern seaboard of the United States with the attendant Low centered over Labrador, Canada. Another Low was centered in the vicinity of Lake Winnipeg, Canada, with its eastern periphery slightly west of Duluth, Minn. This latter Low, although weak, had cold-type characteristics and appeared to be nearly vertical to a minor surface depression. A pronounced belt of westerlies prevailed from the Northern Rockies to and beyond the coastline of Georgia, an unusually far southward displacement of the westerlies over the eastern portion of the country for the month of July.

The portion of the cold front which extended westward from the longitude of Bermuda and thence across the Southeastern States had practically ceased its southward progression by 0030 GMT of the 19th (fig. 3D). As the southward movement diminished, weak waves formed along the front. These waves initially had little significance and produced only minor amounts of precipitation. But there were indications that a wave formation near the junction of the States of Texas, Oklahoma, and Arkansas held considerable promise for future intensification. There, a developing flow of southerly winds in the lower levels of the atmosphere was advecting warm air northward from Mexico and the western Gulf of Mexico. However, the current dew point values were relatively low near the front at this time. High pressure, having increased both in area and intensity, stretched from Hudson Bay southeastward to off the New England coast. The surface center of the cold Low was located slightly west of St. Cloud, Minn., drifting east-southeastward. Precipitation over the eastern portion of the country during the past 24-hours had been quite light, but moderate to locally heavy shower activity was increasing in the trough area that extended from Arkansas northward.

At the 500-mb. level, the 0300 GMT chart on July 19 (fig. 3E) indicated a decreased flow of westerlies across the central Appalachians while the cold Low centered west of Duluth, Minn., had become a "cut-off" Low enclosed by the 18,800-ft. contour with an 18,700-ft. central height value. And the upper trough previously along the eastern seaboard was well off shore. Constant pressure heights over Florida had begun to rise after an easterly trough crossed the southern portion of the peninsula and moved into the Gulf of Mexico. An area of weak cyclonic vorticity had begun to develop over Oklahoma.

By the morning of the 20th the 0030 GMT surface chart (fig. 4A) indicated the apex of the wave over northwestern Alabama was enclosed by the 1014-mb. isobar. Other weak waves appeared to the east and west of this Low. The sea level pressure value of the cold Low, 1012 mb. indicated slight intensification and was located near Moline Ill., enclosed within a broad trough area that covered the Mississippi Valley region and wherein most of the precipitation had been confined during the past 24 hours. But it should be noted that considerable rainfall was beginning to occur ahead of the frontal waves outside of this trough area. High pressure continued to build and bridge across the front toward Bermuda. Concomitantly with the northward intensification of the Mississippi Valley trough and the building of the high pressure off the northeastern coast, the western periphery of the Bermuda High cell moved toward the southeast coast.

At the same time in the upper air the 500-mb. constant pressure chart (fig. 4B) indicated an expanding and increasing cyclonic flow about the cold Low centered southeast of Minneapolis, Minn. In fact, a zone of confluence was developing over Georgia and the Carolinas as the southerly flow over Florida increased from the westward building of the Bermuda anticyclone. With the establishment of the cold Low in the western Great Lakes region there was increased advection of cool outbreaks southward and eastward that produced short-wave troughs. This 500-mb. synoptic chart indicated a pronounced sharpening of the easternmost trough over eastern Georgia and eastern Tennessee, and a less pronounced trough extending from northwestern Missouri to eastern Oklahoma. The belt of westerlies had disappeared as the flow became more meridional due to the building of the ridges over western United States and off the east coast and to a Low centered over the North Central States. This developing southerly meridional flow over the Southeastern States was being intensified by a confluence zone in the area directly north of the sharpening trough. This increased northward advection of the thermal field (see thickness chart, fig. 4A) suggested further intensification of the off-shore ridge along with an increasing zone of maximum anticyclonic vorticity in the thermal field east of the Appalachian Mountains. The area of maximum cyclonic vorticity was in an excellent position to produce intensification of one of the waves.

During the subsequent 36-hour period from 0030 GMT of July 20 to 1230 GMT, July 21, exceedingly heavy rains occurred along a narrow strip extending from the vicinity of Savannah, Ga., northward into southeastern Pennsylvania then curving eastward into portions of extreme southern New England. Further discussion concerning this narrow region of heavy precipitation will be covered in other sections of this paper.

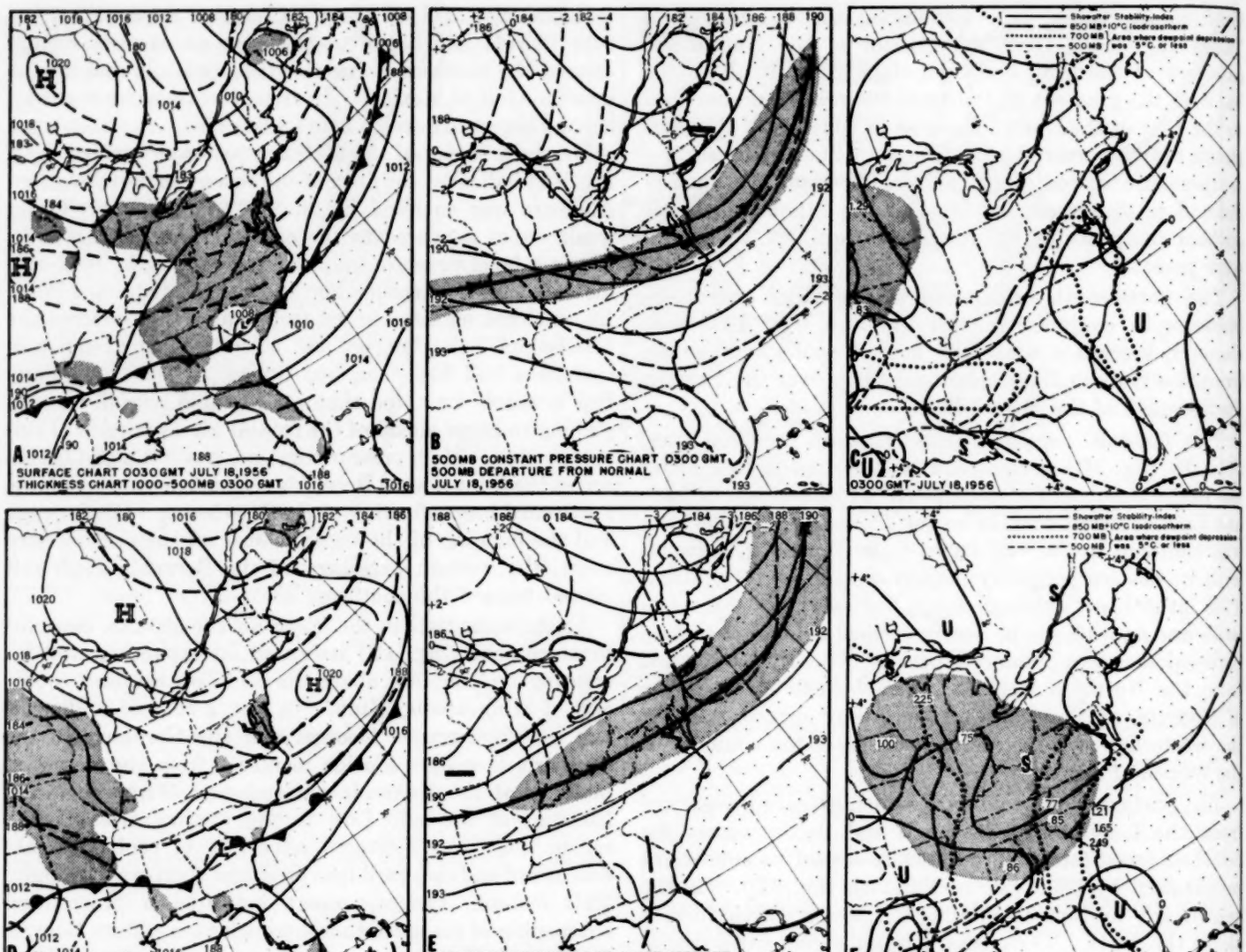
A
SURFACE CHART 0030 GMT JULY 18, 1956
THICKNESS CHART 1000-500 MB 0300 GMT
B
500 MB CONSTANT PRESSURE CHART 0300 GMT
500 MB DEPARTURE FROM NORMAL
JULY 18, 1956
C
0300 GMT - JULY 18, 1956
D
SURFACE CHART 0030 GMT JULY 19, 1956
THICKNESS CHART 1000-500 MB 0300 GMT
E
500 MB CONSTANT PRESSURE CHART 0300 GMT
500 MB DEPARTURE FROM NORMAL
JULY 19, 1956
F
0300 GMT - JULY 19, 1956

FIGURE 3.—(A) Sea level pressure pattern (solid lines) with fronts, 0030 GMT, and 1000–500-mb. thickness pattern (dashed lines) for 0300 GMT, July 18, 1956. Shading shows 24-hour precipitation area ending at 0630 GMT, July 18. (B) 500-mb. contour pattern (solid lines), 500-mb. departure from normal height (dashed lines), 200-mb. jet stream (heavy line with arrows indicating direction of movement), and winds of 75 knots or stronger (shaded area) all for 0300 GMT, July 18, 1956. (C) Composite chart of six semi-related items. 850-mb. dew point areas of +10° C. or higher (within long-dashed lines), 700-mb. areas with dew point depression 5° C. or less (within dotted lines), 500-mb. areas with dew point depression 5° C. or less (within short-dashed lines), calendar-day rainfall totals of 0.75 inch or more plotted at first-order stations, and Showalter stability index values (continuous lines at 4° C. intervals) with center of areas of stability and instability marked "S" and "U" respectively; all for 0300 GMT, July 18, 1956. Also shown is the area of vertical motion in excess of 0.005 m. p. s. (shaded) computed by the Joint Numerical Weather Prediction Unit for 1500 GMT, July 18. (D) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern at 0300 GMT, July 19, 1956. Shading shows 24-hour precipitation area ending at 0630 GMT, July 19. (E) 500-mb. contour pattern and departure from normal height, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 19, 1956. (F) Composite chart showing same elements as (C) for 0300 GMT, July 19, 1956. Vertical motion shown is for 1500 GMT, July 19.

Significant changes occurred during the next 24 hours in the pressure field distribution and the location of the front over the coastal States as illustrated by the surface chart for 0030 GMT, July 21 (fig. 4D). The rapid northward movement of the frontal system was concomitant with the intensification of the surface wave with apex near Washington, D. C. There was continued excellent frontal definition by the thickness pattern with strong negative relative vorticity in the thickness pattern about the apex of the wave, a condition considered favorable for the production of maximum rainfall occurrence (see Cressman [4]). It appears probable that this area of negative relative vorticity was advected nearly northward along the path of heaviest precipitation.

In the upper air at 0300 GMT, July 21 (fig. 4E) the rapid building and intensification of the eastern ridge was easily

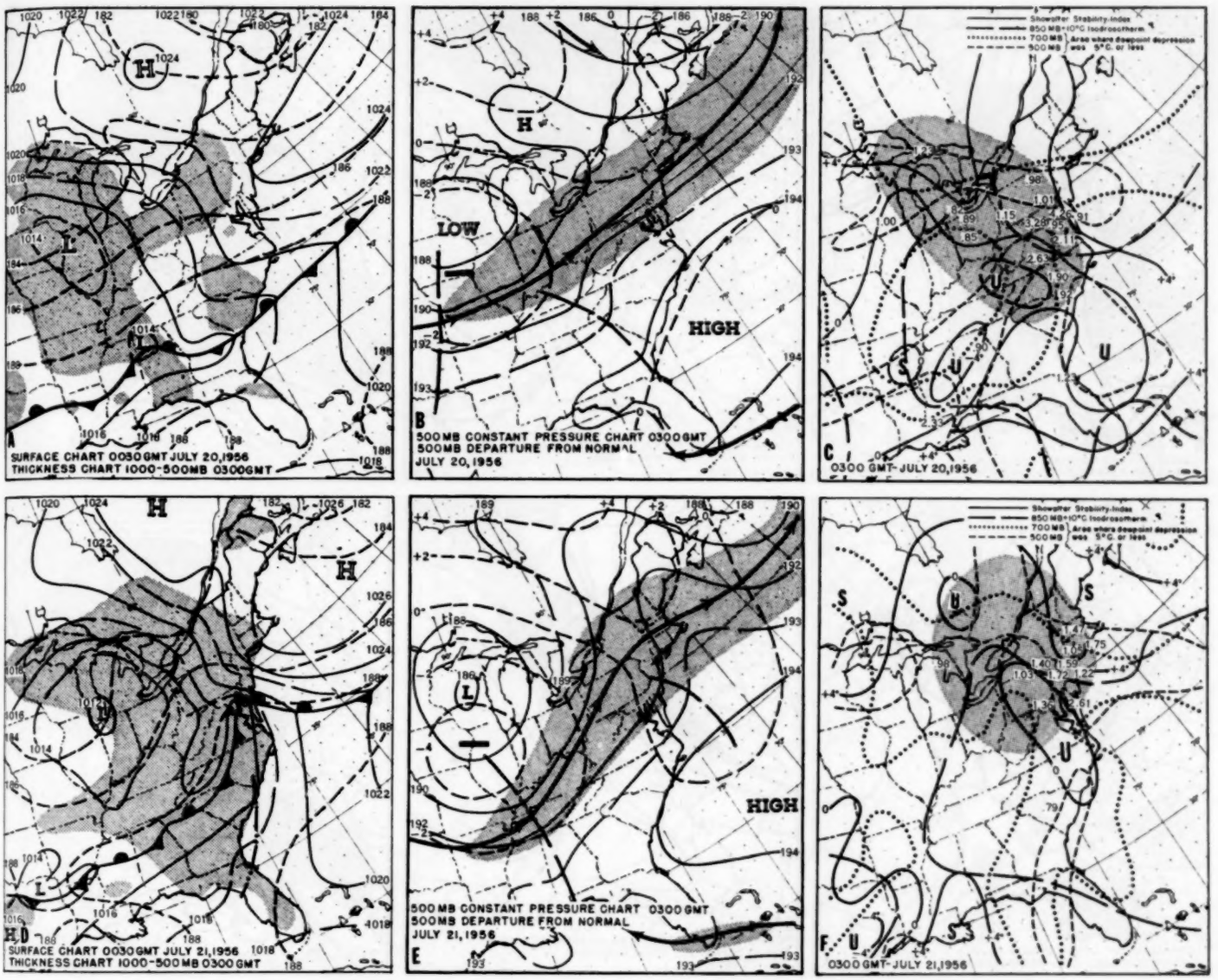
SURFACE CHART 0030 GMT JULY 20,1956
THICKNESS CHART 1000-500MB 0300GMT
500MB CONSTANT PRESSURE CHART 0300 GMT
500MB DEPARTURE FROM NORMAL
JULY 20, 1956
0300 GMT- JULY 20,1956
SURFACE CHART 0030 GMT JULY 21,1956
THICKNESS CHART 1000-500 MB 0300 GMT
500 MB CONSTANT PRESSURE CHART 0300 GMT
500MB DEPARTURE FROM NORMAL
JULY 21,1956
0300 GMT- JULY 21,1956
HIGH
LOW

FIGURE 4.—(A) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern (dashed) for 0300 GMT, July 20, 1956. Shading shows 24-hour precipitation areas ending at 0630 GMT, July 20, 1956. (B) 500-mb. contours and departure from normal, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 20, 1956. (C) Composite chart showing six semi-related items. 850-mb. dew point areas of +10° C. or higher, 700-mb. and 500-mb. areas with dew point depression 5° C. or less, calendar-day rainfall totals of 0.75 inch or more, and Showalter stability index values; all for 0300 GMT, July 20, 1956. Areas of positive vertical motion in excess of 0.005 m. p. s. (shaded) are for 1500 GMT, July 20. (D) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern for 0300 GMT, July 21, 1956. Shading shows 24-hour precipitation area ending at 0630 GMT, July 21. (E) 500-mb. contours and departure from normal, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 21, 1956. (F) Composite chart showing same elements as (C) for 0300 GMT, July 21, 1956. Vertical motion areas for 1500 GMT, July 21.

apparent and with this building of high pressure, a definite block had been established that would hinder any immediate eastward displacement of the cold Low.

Generally over the remaining area of the Eastern States little significant change had occurred at either surface or aloft. Thus, there continued a more or less repetitious occurrence of weak short waves or cool outbreaks from the cold Low, waves along the front (mostly minor), and a slow eastward progression of the ridge area as the cold Low drifted northeastward. These conditions prevailed through the 0030 GMT surface chart and the 0300 GMT upper air charts of the 23d. Discussion concerning these charts has been omitted for the sake of brevity but the 24-hour sequence of charts is uninterrupted in figure 5 A, B, D, E. Precipitation continued to be rather widespread during this period with scattered totals of one or more inches.

On the morning of the 24th (GMT) (fig. 6A, B) the cold Low had again become approximately vertical and about stationary in the vicinity of Moosonee, Ont. Intensification of this Low had occurred at the surface and aloft with a definite trough indicated at upper levels from the

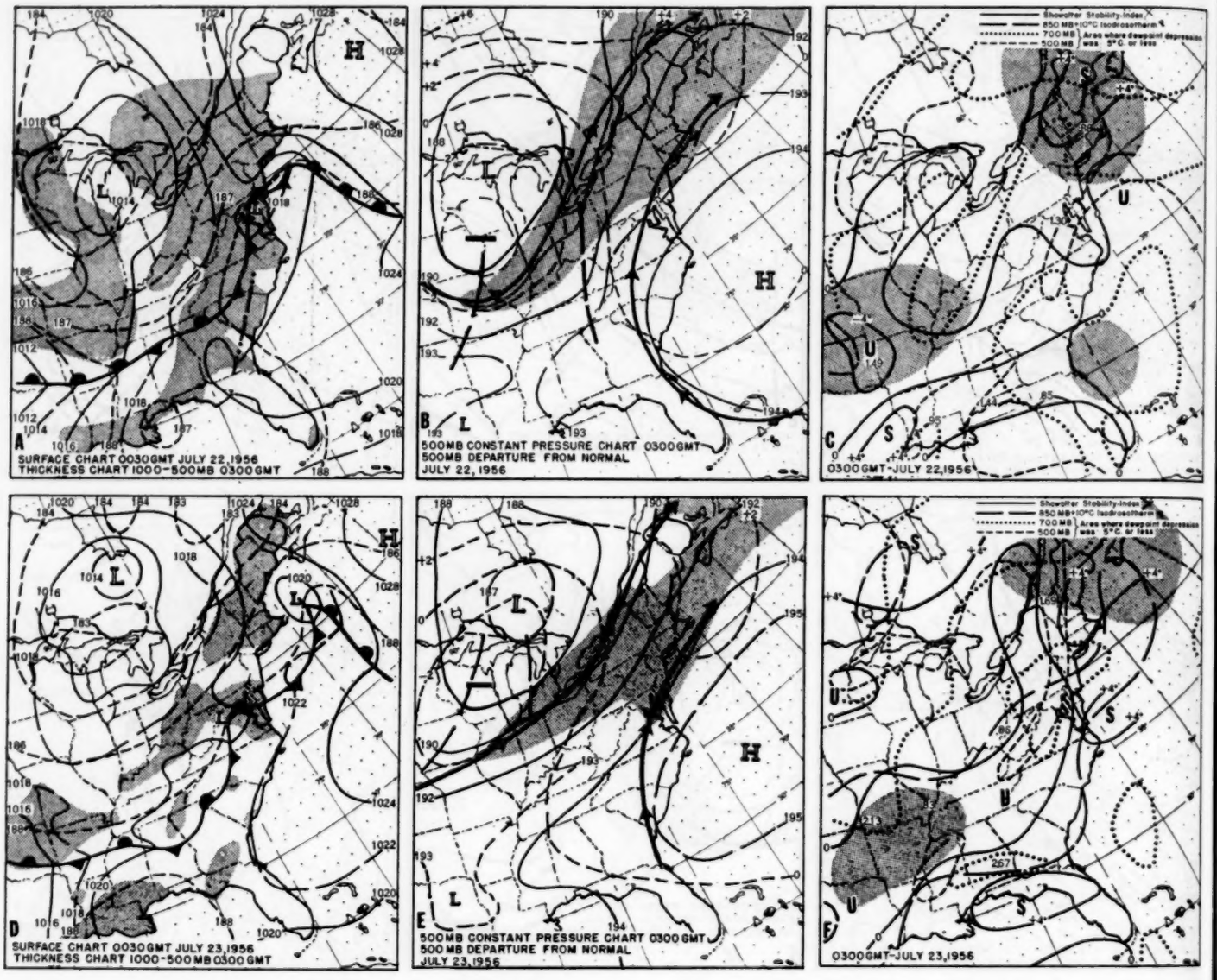

FIGURE 5.—(A) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern (dashed) for 0300 GMT, July 22, 1956. Shading shows 24-hour precipitation areas ending at 0630 GMT, July 22. (B) 500-mb. contours and departure from normal, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 22, 1956. (C) Composite chart showing six semi-related items. 850-mb. areas of +10° C. or higher, 700-mb. and 500-mb. areas with dew point depression 5° C. or less, calendar-day rainfall totals of 0.75 inch or more, and Showalter stability index values; all for 0300 GMT, July 22, 1956. Areas of positive vertical motion in excess of 0.005 m. p. s. (shaded) are for 1500 GMT, July 22. (D) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern for 0300 GMT, July 23, 1956. Shading shows 24-hour precipitation area ending at 0630 GMT, July 23. (E) 500-mb. contours and departure from normal, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 23, 1956. (F) Composite chart showing same elements as (C) for 0300 GMT, July 23, 1956. Vertical motion areas for 1500 GMT, July 23.

upper Great Lakes to northeastern Texas. Over the Atlantic there had been a gradual weakening and continued eastward drift at all levels of the anticyclone. To the rear of this High an occluded front and attendant Low, which previously was the pronounced wave near Washington, D. C., were headed toward Sable Island. In conjunction with these features was the return toward a more zonal flow over the Eastern States.

During the next 24-hour period (fig. 6D) there was pronounced weakening and partial dissipation of the frontal system over the eastern half of the nation in conjunction with decreasing thermal gradient. Concordant with this proceeding was the rapid invasion of another front over the upper Great Lakes region; the thermal pattern defining this front indicated that it had more of the characteristics of an occlusion or a trough of warm air aloft (TROWAL, as used by the Canadian Meteorological Service [5]) rather than a cold front. Pressure ahead of this front was falling as the High cell over the Atlantic Ocean continued to withdraw eastward. Precipitation persisted over a wide area to the north and west of the deteriorating front as a new zone of rainfall appeared in the upper Great Lakes.

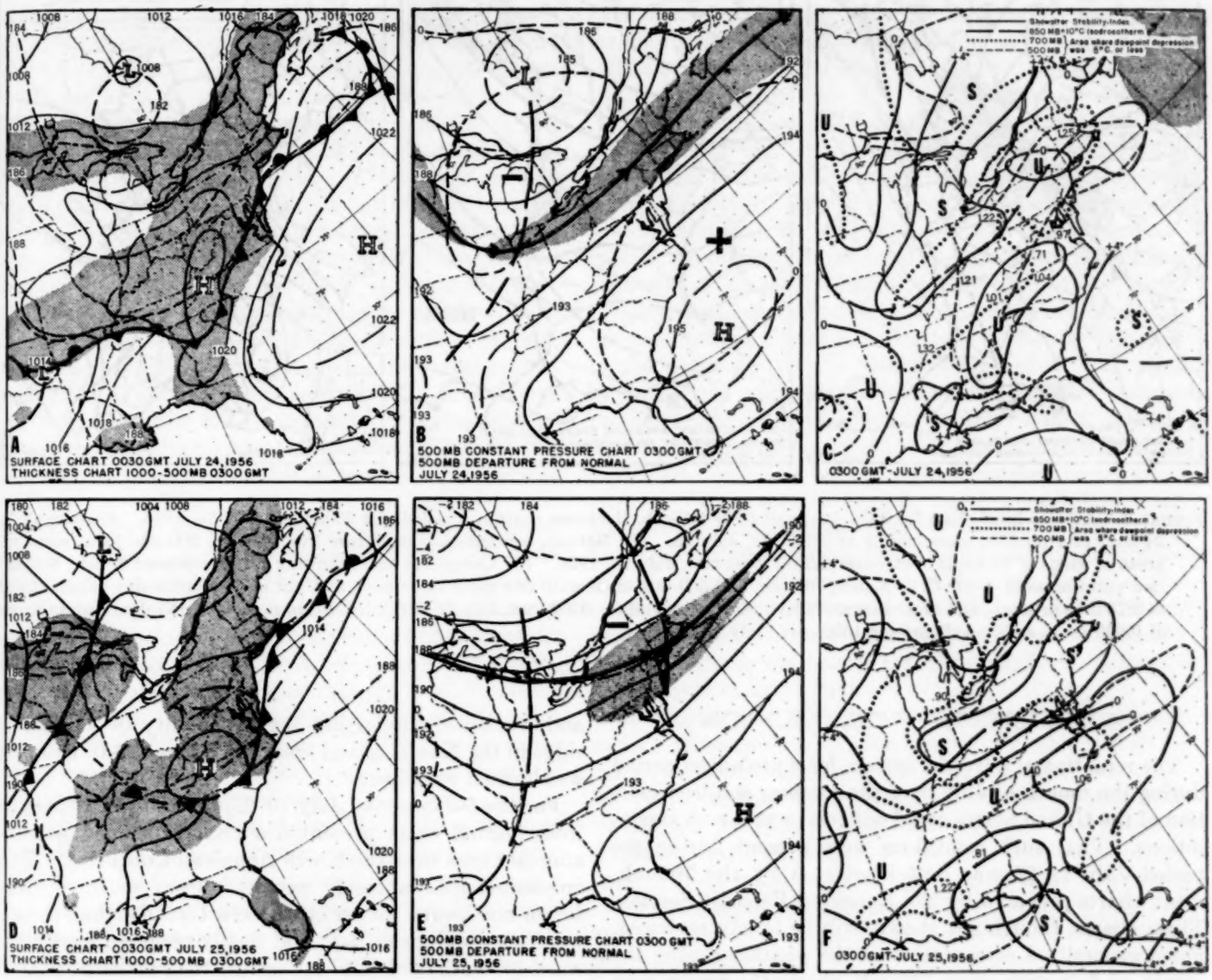

FIGURE 6.—(A) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern (dashed) for 0300 GMT, July 24, 1956. Shading shows 24-hour precipitation areas ending at 0630 GMT, July 24. (B) 500-mb. contours and departure from normal, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 24, 1956. (C) Composite chart showing six semi-related items. 850-mb. dew point areas of +10° C. or higher, 700-mb. and 500-mb. areas with dew point depression 5° C. or less, calendar-day rainfall totals of 0.75 inch or more, and Showalter stability index values; all for 0300 GMT, July 24, 1956. (D) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern for 0300 GMT, July 25, 1956. Shading shows 24-hour precipitation area ending at 0630 GMT, July 25. (E) 500-mb. contours and departure from normal, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 25, 1956. (F) Composite chart showing same elements as (C) for 0300 GMT, July 25, 1956. Vertical motion areas for 1500 GMT, July 25.

In harmony with these surface developments the 500-mb. level (fig. 6E) indicated a continued zonal flow over the central and northern portions of the eastern United States while the upper trough that had been somewhat persistent over the western portion of this area was being displaced eastward ahead of the new front.

As the closing phase of this situation on July 26 (fig. 7A, B), frontal decay removed the old discontinuity from the Eastern States and weakening occurred along the remaining portion of the front south of the 40th parallel in the eastern coastal waters. The new front continued its rapid eastward advance in the Great Lakes area. Approximately midway between these fronts was a rather well-defined trough apparent at the surface as well as aloft. It appeared that this surface trough was the reflection of the old upper trough that had been displaced rapidly eastward during the past 24 hours. Numerous showers and thunderstorms attended the passage of this trough while to its rear there was a general decrease in moisture and cloudiness, and a shift of the gradient winds to a more westerly direction. Thus a prolonged period of cool, damp weather from a spring-like frontal wave pattern was brought to a close.

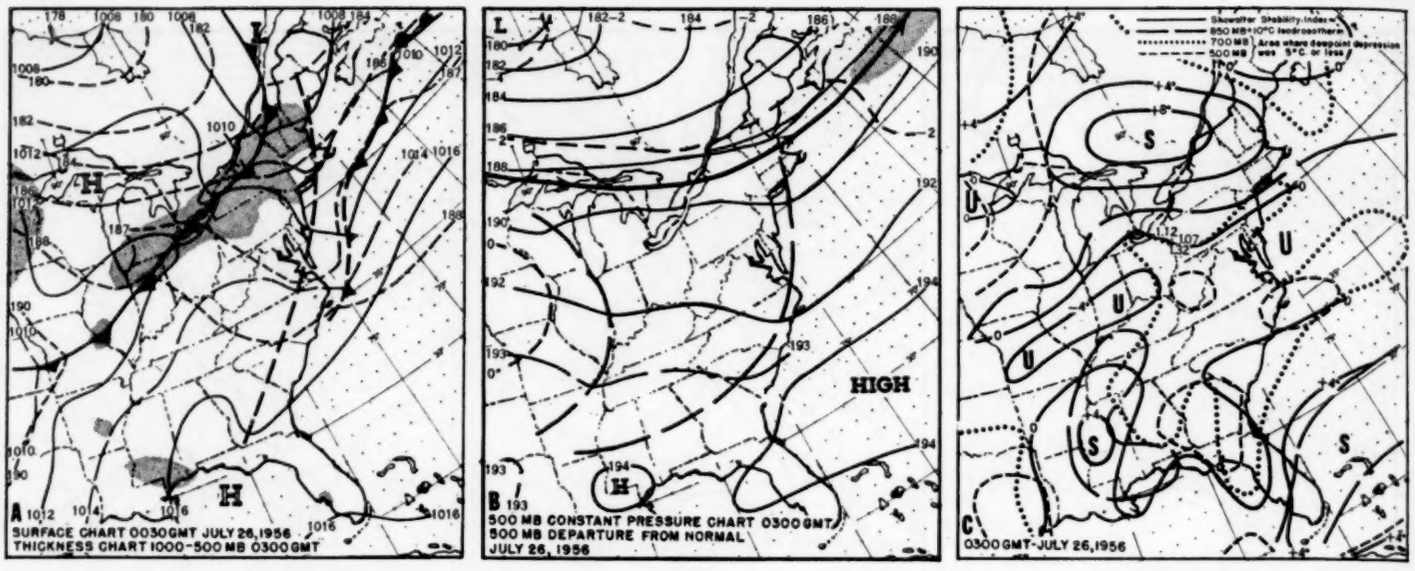

FIGURE 7.—(A) Sea level chart for 0030 GMT, and 1000–500-mb. thickness pattern (dashed) for 0300 GMT, July 26, 1956. Shading shows 24-hour precipitation areas ending at 0630 GMT, July 26. (B) 500-mb. contours and departure from normal, 200-mb. jet stream, and areas of winds of 75 knots or stronger; all for 0300 GMT, July 26, 1956. (C) Composite chart showing six semi-related items. 850-mb. dew point areas of +10° C. or higher, 700-mb. and 500-mb. areas with dew point depression 5° C. or less, calendar-day rainfall totals of 0.75 inch or more, and Showalter stability index values; all for 0300 GMT, July 26, 1956. Areas of positive vertical motion in excess of 0.005 m. p. s. (shaded) are for 1500 GMT, July 26.

4. PERSISTENCY OF THE COLD LOW

The retention of a frontal system for a prolonged period during the month of July over the extreme southern portion of the United States could only occur under ideal conditions. That such conditions were present during the period July 18–25 was best illustrated by the 700-mb. mean charts prepared by the Extended Forecast Section. These charts, though not shown, will be briefly reviewed. For a generalized presentation of the weather during the month of July with mean 700-mb. charts the reader may refer to the article in this issue by Krueger [15].

The 700-mb. departure from normal chart for the Northern Hemisphere for the 5-day period July 14–18 inclusive, depicts an area of strong positive anomalies encompassing the northern latitudes except for an intense negative anomaly of considerable size located between the North Pole and Alaska. From this hemispherical ring of positive anomalies, an arm-like area stretched southward along the western coast of North America reaching inland over the Southwestern States. Concurrently a large negative anomaly area existed over the New England States.

By the period of July 17–21 inclusive, the 700-mb. 5-day departure from normal chart presented the positive anomaly area over primarily the same region, although some intensification had occurred along the west coast of North America with a 320-ft. positive anomaly area over the northern portion of Vancouver Island and a plus 380-ft. anomaly center over northwestern Hudson Bay. The negative area over the New England region had split and weakened, indicating the probability of a building ridge in the New England area. A minus 130-ft. anomaly was located over Iowa.

For the 5-day period July 19–23, the 700-mb. departure from normal chart presented a crescent-shaped positive anomaly area over the North American Continent. This crescent-shaped anomaly extended from southern California northward into southwestern Canada, thence eastward across Hudson Bay, and finally recurved southeastward to southward until it reached the northern coast of Georgia. This positive anomaly area attained a value of 460 ft. over the central portion of Hudson Bay. In this position it completely blocked any movement of the negative departure from normal area that had a central value of minus 110 ft. over eastern Iowa.

This blocking condition was maintained in much the same pattern on the next 5-day 700-mb. departure from normal chart, except there had been a slight northeastward drift of the center of positive anomaly previously over the central portion of Hudson Bay thus allowing the below normal center with central value of minus 120 ft. to drift eastward into the southern portion of Lake Michigan.

5. TEMPERATURE DISTRIBUTION

The arrangement of the cyclonic and anticyclonic pressure areas over the central and eastern portions of Canada presented an ideal synoptic model for a wide-scale flow of cool air to stream southward from the Hudson Bay region into the northeastern sector of the United States. Since the Hudson Bay region is one of the main source areas

for cool air during the summer months on the North American Continent, and since the northeastern section of the United States was experiencing a period of cool weather, we have an ideal situation for the advent and occurrence of much below normal temperatures. The outstanding temperature features of this study were not of record-breaking low readings at any one station or any group of stations, but rather the duration of these relatively pronounced departures from normal temperatures over a large area, i.e., most of northeastern United States. As shown by figure 1, the average daily temperatures were below normal every day from July 18–25 inclusive from the Ohio Valley to east-central Pennsylvania. In fact, the entire area north of the front was below normal at least 75 percent of the time. The area to the lee of the lower Great Lakes had the lowest frequency of below normal average temperatures in the Northeastern States as one would expect, nevertheless this area was below normal throughout most of the period.

In the *Weekly Weather and Crop Bulletin* [6], the chart depicting the departure of average temperatures from normal for the period July 16–22 inclusive (not shown) illustrates that all of the eastern half of the United States was below normal with the exception of the Southern Coastal States from Texas to North Carolina inclusive. The greatest temperature departures occurred from the Northern Plains through the New England area with anomalies of −6° F. or more, and locally in the upper Great Lakes region the departures were in excess of −9° F. during that 7-day period.

Further investigation has shown that on one or more days during the period of July 18–25, the maximum temperature receded to values of 10° F. or more below the monthly normal in the States north of the 37th parallel (fig. 1). This figure also indicates that at some time during the period nearly all of the eastern half of the nation recorded afternoon high temperatures 5° F. or more below normal monthly values, and that a small area in Michigan reported a departure of −20° F. with an almost equal value reported later in southwestern Missouri. It was apparent these low maximum temperatures were the principal contributing factor to the below normal average temperature that prevailed in the eastern half of the country. For, in contrast to the magnitude of the departure from normal of the maximum temperatures, the minimum readings were more on the order of 6° to 13° F. below the monthly minimum value over the Northeastern States, and were 3° to 7° F. under the normal values over the Southeastern States.

If one were to draw isochrones for the date of lowest maximum temperature during the period of July 18–25, this chart (not shown) would illustrate that the cold air had invaded only the northern portions of Illinois, Indiana, Ohio, and the extreme western portions of New York and Pennsylvania on July 18. But in the following three days this cooling engulfed most of the Northeastern States and a goodly portion of the South. However, over many of the Southern States the coldest daily maximum temperature did not occur until July 23–24.

The comparatively shallow cold air that prevailed over the Northeastern States due to the existent wind flow pattern could not possibly have been responsible for this late southern cooling, thus another source region must be located. Close examination of the 500-mb. temperatures as early as July 20 shows there was a marked thermal trough from western Kansas to the Texas Panhandle. As this cold thermal trough moved into the central and lower Mississippi Valley the 5- and 6-km. temperatures equaled and in some cases exceeded the minimum July temperature extremes previously recorded in the upper air [3]. It was this area of cold air that produced the rapid development of positive vertical motion near Springfield, Mo., on July 22. This comparatively deep mass of cold air served as the second source area for below normal surface temperatures. Concurrently with this happening the low-level flow over the Northern States east of the Appalachian Mountains had reversed and had become a more easterly or southeasterly flow. The source region for this flow was the airmass then over the region from Hudson Bay to Nova Scotia. This east-to-southeast flow of air was sufficiently cool to continue the below normal temperatures throughout the Northeastern States.

In conjunction with our discussion of below normal surface temperatures for this period it would be well to make reference to the departure from normal of the 1000–500-mb. thickness charts (not shown). In general, over the eastern portion of North America these charts, while similar in their general appearance to the 500-mb. departure from normal chart (figs. 3 to 7, B and E), were not quite as encompassing in area or intensity of negative values. Generally the extreme negative departures from the normal thickness values for July were under 400 ft. below the norm. For general comparison it might be stated that the 1000–500-mb. thickness pattern during this time was in rather good agreement with the 1000–500-mb. normal thickness for the month of September.

6. MOISTURE IN THE UPPER AIR

As previously mentioned there was no wide-scale transport of moisture directly northward from the Gulf of Mexico, July 18–25. The moisture that did invade the United States appeared to have traveled a rather circuitous route, either from the eastern Pacific Ocean or northward from the Gulf of Mexico over eastern Texas. However, in the latter days of this period the transport of moisture from the southern waters became somewhat more direct under a weak anticyclonic flow.

For the location of moisture areas at 0300 GMT, July 18, and at 24-hour synoptic periods thereafter, at the constant pressure levels of 850, 700, and 500 mb., a series of charts were prepared (figs. 3 to 7, C and F). On these charts has been reproduced by long-dashed lines the area within which at 850 mb. all dew points were 10° C. or higher.

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By means of dotted lines regions of moisture at 700 mb. have been depicted; within these enclosed areas the dew point departure from the observed temperature was 5° C. or less. The short-dashed lines represent the moisture areas at 500 mb. using the same definition as was applicable to the 700-mb. level. The numerical values of two or more digits scattered over the diagram represent calendar-day rainfall totals of 0.75 inch or more which occurred at a nearby first-order station.

With the inclusion of these moisture areas on the charts and accounting for normal advection for each 24 hours with attendant growth or shrinkage of these regions of moisture, it was observed by comparison (figs. 3 to 7, C and F) that practically all of any previous 24-hour rainfall pattern was enclosed by two or more of the moisture areas. Furthermore, it was clearly indicated that wherever heavy precipitation occurred these three levels of moisture were, for all practical purposes, superimposed over that area. In addition, it was interesting to note that the lack of moisture at the 700- and 500-mb. levels rather clearly differentiated the areas of no rainfall even though the 850-mb. dew point values were greater than 10° C.

Finally it was decided to include a fourth index upon these moisture charts and so, by use of continuous lines, the positive and negative areas of stability, as derived by Showalter [7], have been indicated.

7. INSTABILITY

With the interaction of the cold Low and its attendant low temperatures in the cP airmass to the north of the front, in direct contrast to the warm moist mT airmass south of the front, there is little doubt as to the probable occurrence of considerable instability. This instability was further indicated during the period by the development of showers and thunderstorms, as well as a few tornadoes. The Showalter Stability Index Charts [7] as prepared by the National Weather Analysis Center and presented in figures 3 to 7, C and F, show more quantitatively the distribution of the areas of convective stability and instability. These areas are outlined by the solid continuous lines with the positive areas marked by a plus sign and the negative area by a minus sign. Even a cursory examination of these charts clearly indicates the large negative or unstable zones near the fronts and in the areas associated with large rainfall totals. It must be remembered that the area of instability on these charts is for the time 0300 GMT and normal movement along the wind-flow pattern must be expected during the next 24-hour period in accounting for precipitation patterns that might be indicated as outside of its present area.

8. VERTICAL MOTION

For the areas where vertical motion prevailed during this period we shall rely on the charts that were prepared by the Joint Numerical Weather Prediction Unit (JNWP). The thermotropic model, utilizing the 1000-mb. and 500-mb. levels and dry adiabatic processes, yields an average vertical velocity which can be assumed to be centered near the 750-mb. level. Furthermore, while these velocities are representative of the average velocity over the area between grid points, it does not mean that smaller areas of increased or decreased vertical velocities could not occur within these smoothed contours. However, it was thought that it would be of interest to compare the location and intensity of the areas of positive vertical velocities, as produced by machine, with regions where precipitation occurred. It must be remembered that these vertical motion charts are prepared at 1500 GMT only and are 12 hours out of phase with all other upper air data that are shown. The shaded area on figures 3 to 7, C and F was the region where positive vertical velocity of 0.005 m. p. s. or higher was computed by the machine. The printed figures scattered over these charts are rainfall totals of 0.75 inch or more that occurred during the calendar day at a first-order weather station. Or, comparison may be made with the 24-hour precipitation pattern as depicted by the shaded area (figs. 3 to 7, A and D).

In order not to complicate figures 3 to 7, C and F further, the central values within the area of positive vertical motion were omitted, but for those who may be interested in these values they are as follows: July 18, 0.01 m. p. s.; July 19, 0.015 m. p. s.; July 20, 0.017; July 21, 0.015; July 22, 0.012, plus a new development indicated by a positive area that was formed by strong barotropic effects near Springfield, Mo., with 0.016 m. p. s. at the center (the day previous the highest positive velocity within reasonable radius was a 0.002 m. p. s. value); July 23, 0.008 m. p. s. near Eastport, Maine, and 0.007 m. p. s. near Little Rock, Ark.; July 24, 0.016 m. p. s. southeast of Sable Island, while the area previously over Arkansas dissipated.

It is interesting to note that while this area of vertical motion encloses a goodly portion of the rainfall area, it again leaves, as have other factors discussed in this paper, the utopia of precipitation forecasting far removed.

For example, by the use of Showalter's formula [8] in the computation of rates of precipitation expectancy from vertical motion:

$$I=\frac{w_0\rho_0(x_0-x_1)}{7}$$

where I is precipitation in inches per hour; w_0 is the vertical velocity in meters per second at the bottom of the column, i. e., condensation level; ρ_0 is the density of the air in gm./m.^{3} at the condensation level, and (x_0-x_1) is the difference in mixing ratio in gm./gm. at the condensation level and at the top of cloud or top of lift.

This formula has been further reduced by Showalter [9], using an approximation method, and designed for ease and speed of computation rather than for rigid

exactness during the rapid preparation of quantitative rainfall estimations. He developed the following approximations dependent upon the surface dew point values:

For a dew point of	*I equals*
75°F.	$2w$
50°	w
35°	$w/2$

I=inches of rainfall per hour
w=vertical motion in meters per second

In cases such as warm frontal rains or orographic rainfall where slope is known, w can quickly be figured for the lift since it is a product of the slope and the speed of the wind component normal to the slope.

Furthermore, this abbreviated formula furnishes an easy and rapid means of computing the approximate value of the vertical velocity when the amount of precipitation that has occurred in a given length of time is known. For example: At Baltimore, Md., on July 20 it was reported that 4.26 inches of rain occurred in 24 hours, of which 4.25 inches fell in the last 12 hours of the day, 3.24 inches were collected in a 4-hour period, and 1.35 inches in a 1-hour period. The surface dew point temperature was near 72°F. Using the approximation formula and a value of $1.9w$ for the dew point of 72° F., we find that the vertical velocities were approximately 0.20 m. p. s. for the 12-hour period, 0.50 m. p. s. during the 4-hour rainfall, and 0.70 m. p. s. during the 1-hour downpour.

9. JET STREAM ANALYSIS

In this study it is interesting to note that the occurrence of a low-level jet stream was missing for the rapid transport of moisture from the Gulf of Mexico into the areas of heavy precipitation. Thus it appears that the bands of moisture necessary for the production of heavy rains over the Southeastern States arrived in these areas by means of normal advection, consequently eliminating the rapid replacement of moisture into these storm areas as would have occurred under low-level jet stream conditions. This in all probability reduced the total amount of precipitation that otherwise might have been obtained.

Generally the low-level winds over the Eastern States that were located north of the front and beneath the 7,000-ft. level maintained a cyclonic flow and a speed of 10–20 knots throughout the greater portion of the period. South of the front and beneath the same height level winds were variable as to direction while speeds were usually from 7 to 15 knots but locally approached 12–25 knots over Texas. Thus it appears that we must turn our attention to the upper levels of the atmosphere for a jet stream that had any influence upon the occurrence of precipitation.

The 300-mb., 200-mb., and 150-mb. charts all indicated that a well-defined jet stream was present on July 18 (fig. 3B) and was in good correlation with the surface front as observed by Palmén and Newton [10]. At these levels the jet stream continued to be maintained and remained in good agreement with the surface frontal position throughout most of the period. During July 18–21 there was also an easterly jet stream about the southern portion of the Bermuda High. This easterly flow of air apparently contained but little moisture since radiosonde observations reported that "motor-boating" had occurred in the lower levels of this jet stream.

It may be noted from the area of precipitation shown by the shaded zone on each chart (figs. 3 to 7, A and D) and by the position of the 200-mb. jet stream as illustrated on the 500-mb. constant level chart (figs. 3 to 7, B and E) that a large area of rainfall occurred during the period from July 18–25 along and to the north of the jet stream, although an almost equal-sized zone was reported to the south of the jet. In fact on the 20th the heaviest period of precipitation over a moderately sized area occurred well to the south of the jet stream which is in contrast to the findings of Starrett [11] and also of Riehl [12] who have stated that the concentration of precipitation both in frequency and amount usually occurs in or slightly to the north of the jet stream. However, it does appear in this case that the relation between frequency of rainfall and the location of the jet stream might have had a much closer agreement (see jet stream positions, figures 3 to 7, B and E and figure 2).

The southward displacement of the westerlies during this period is illustrated in figure 2, where the average position of the 700-mb. jet stream for July 1956 is shown to be about 300 miles south of the normal position for July jet streams. The actual average position of the jet stream for July 1956 is in relatively close agreement over the eastern half of the Nation with the normal positions of the jets for the month of May or November.

10. TROPOPAUSE RELATIONSHIP

In conjunction with the investigation of the high-level isotach charts the tropopause chart was studied in regard to possible breaklines in the regions of the strong jet streams at the 300-, 200-, and 150-mb. levels. It was thought that there might be some correlation between the area of heavy precipitation over the region from the Carolinas to southeastern Pennsylvania on the 19th and 20th. This was occasioned by the recent study by Culkowski [13] who mentioned that the southern or eastern edges of a tropopause breakline were the optimum locations for heavy precipitation. However, he cautioned that during the summer months the tropopause break area diminished in value as an indicator of excessive precipitation. In this case it appears there was no apparent correlation between the tropopause and the surface precipitation.

11. REGION OF HEAVIEST PRECIPITATION

It so happened during this period, July 18–25, that the area of maximum precipitation and the zone of greatest intensity of short-duration rainfall were in excellent con-

formity with each other. In reality this zone of excessive precipitation was a narrow band approximately 60 miles in width at its broadest part, which happened to be near Washington, D. C., and then tapered to blunted points near its northern and southern limits in the vicinities of Philadelphia, Pa., and a short distance south of Florence, S. C.

A combination of sundry factors conduced this unusual and heavy deposit of rain; in summation these contributing elements included:

1. An active front over the Southern States with stable waves traveling along this front.
2. A cold Low centered near Chicago, Ill., transporting southward some of the coolest air reported for July at 500 mb. and producing recurring impulses of cold air that aided in the genesis of waves.
3. An area of pronounced cyclonic vorticity at the 500-mb. level in phase with a developing and moving surface wave which progressed northeastward toward Washington, D. C., and which had begun acquiring unstable characteristics.
4. An intensification of a high pressure block both at the surface and aloft directly ahead of the developing wave.
5. A rapid sharpening of the upper trough attending the developing wave as it approached the building ridge that was holding over New England and along the eastern seaboard.
6. Creation of an area of pronounced cyclonic curvature and convergence by the intensification of the trough, a normally necessary feature according to Oliver and Oliver [14].
7. Advection into this region of 13° to 14° C. dew points at the 850-mb. level.
8. Advection of near saturated air over the region at the 700-mb. and 500-mb. levels.
9. A zone of positive vertical motion moving across the northern portion of this area.

Thus it would appear that conditions were nearly ideal for the occurrence of excessive rainfall.

Another condition or indication that appeared favorable for the occurrence of heavier rainfall along this narrow strip was the developing area of maximum anticyclonic vorticity displayed by the 1000–500-mb. thickness pattern just to the west of the region at 0300 GMT on the 20th; such an area is considered to be a region of maximum rainfall occurrence (see [12]). During the next 24 hours the thickness pattern developed an area of exceedingly strong anticyclonic vorticity and for the next 24 hours this zone continued to be advected across the southern portion of the New England States but with a considerable decrease in the negative relative vorticity.

12. PARTIAL EFFECTS FROM STORMS

During July 18–25, thunderstorms were frequent over a widespread area of the East; lightning killed several persons and damaged some property. Tornadoes were reported at Lorain, Ohio, Little Muskego Lake, Wis., Pennock, Minn., Falmouth, Ky., and Richmond, Va. Several severe windstorms were recorded. Hail in some local areas did considerable damage to crops and produce; Falmouth, Ky., reported 3 to 4 inches of hail on the ground on July 19. Cool, damp weather hindered agricultural interests by retarding crops and delaying the harvest in many areas.

However, heavy rains and the attendant local flooding of numerous streams accounted for the greatest number of lives lost and the largest property damage. Two lives were lost in the rain-swollen and rampant streams near Jackson Springs, N. C., on the 19th after a downpour of 9.15 inches. In Montgomery County, Md., the northern suburban section of Washington, D. C., 7 persons were drowned, 6 in cars driven into or swept into the swollen flash-flooded streams; the seventh was an attempted rescuer. It has been estimated by the River Services Section[1] that in a small core area approximately 20 miles north-northwest of Washington, D. C. at least 9 inches of rain occurred during a period of less than 12 hours and the greater portion of this total probably fell in less than 4 hours. This region is in the headwaters of Rock Creek basin, a moderate-sized creek which flows through the northern half of the District of Columbia. Damage in the Rock Creek basin was quite heavy. Flooding of streams occurred also at numerous places in eastern Maryland. Estimated damages to roads and bridges in southern Maryland totaled $300,000, with no report received from the northern section.

The Farm Bureau furnished a preliminary estimate of damages to farms, livestock, and feed of from $500,000 to $600,000. Thus, for a storm of short duration and of little moisture value, the cost was high.

ACKNOWLEDGMENTS

The writers wish to express their appreciation to these various units of the U. S. Weather Bureau: River Services Section for information furnished, staff members of NWAC for helpful suggestions and reviewing of the article, and Daily Map Unit for detailed drafting of the figures.

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[1] The information received from River Services Section is tentative and may be subject to slight modification.

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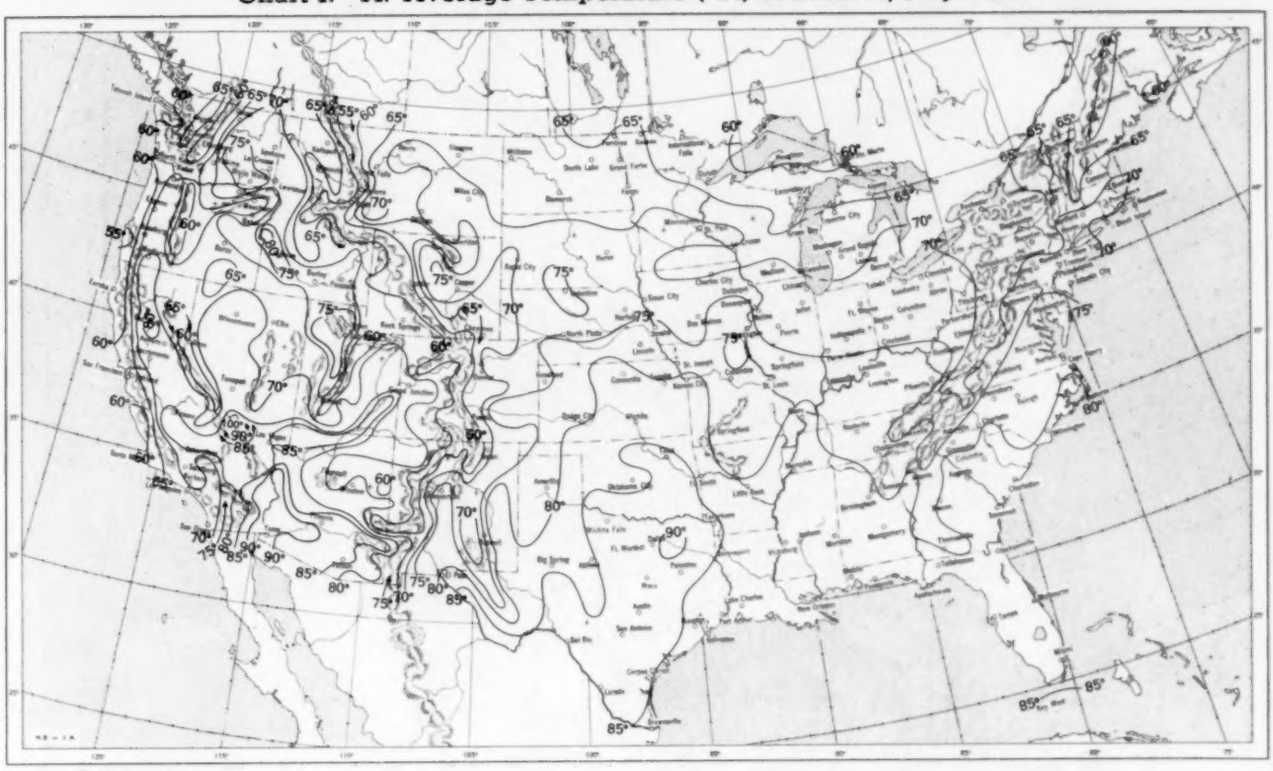
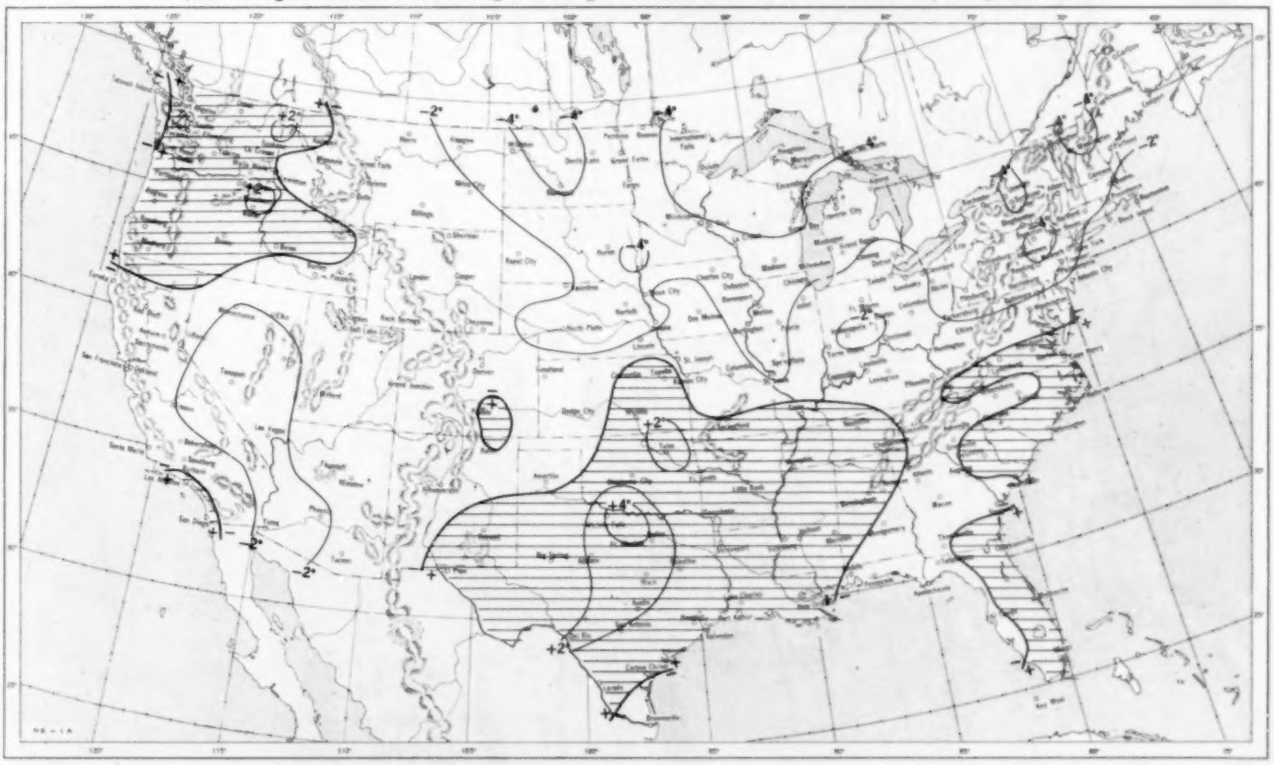
Chart I. A. Average Temperature (°F.) at Surface, July 1956.

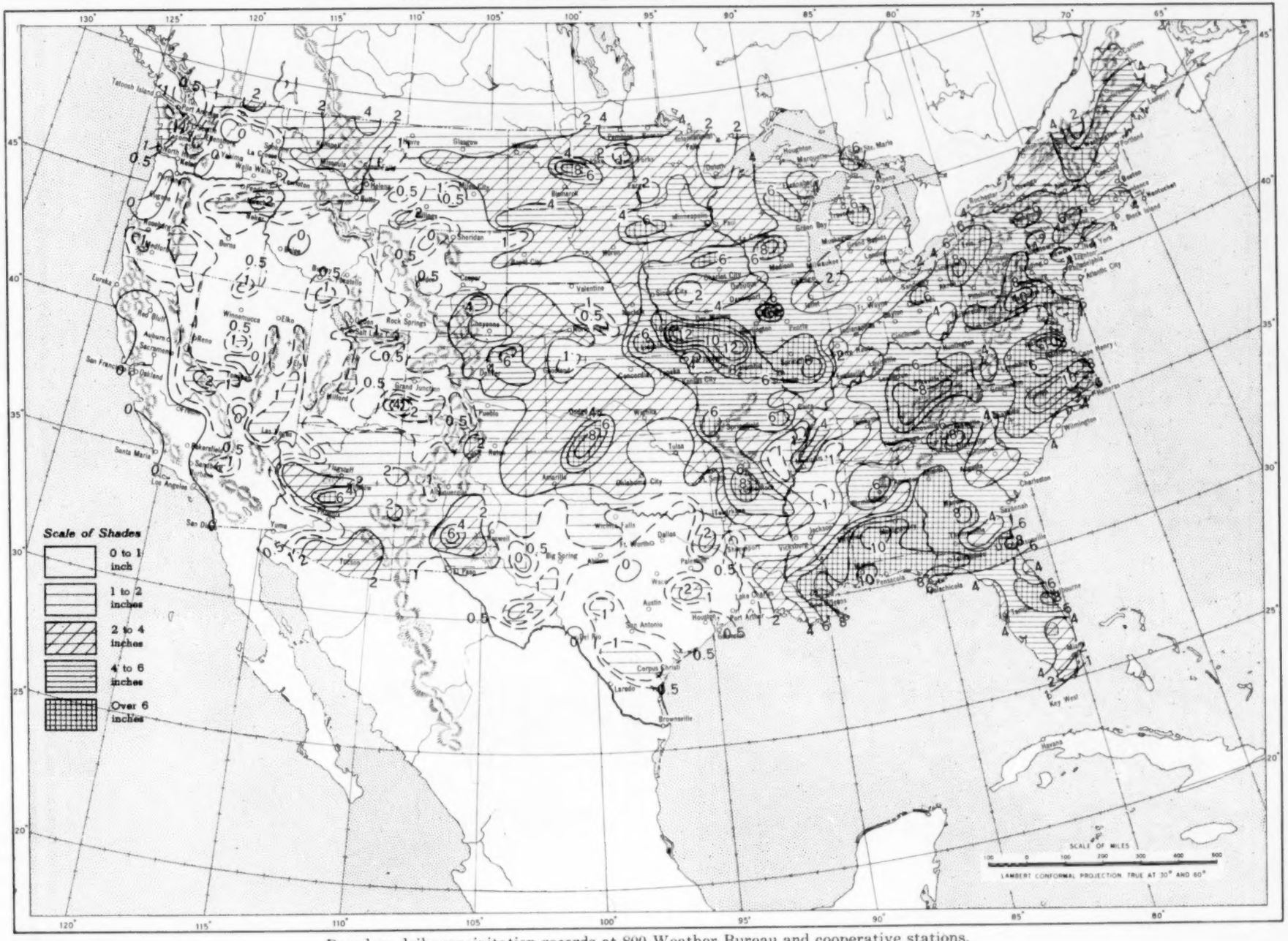


B. Departure of Average Temperature from Normal (°F.), July 1956.



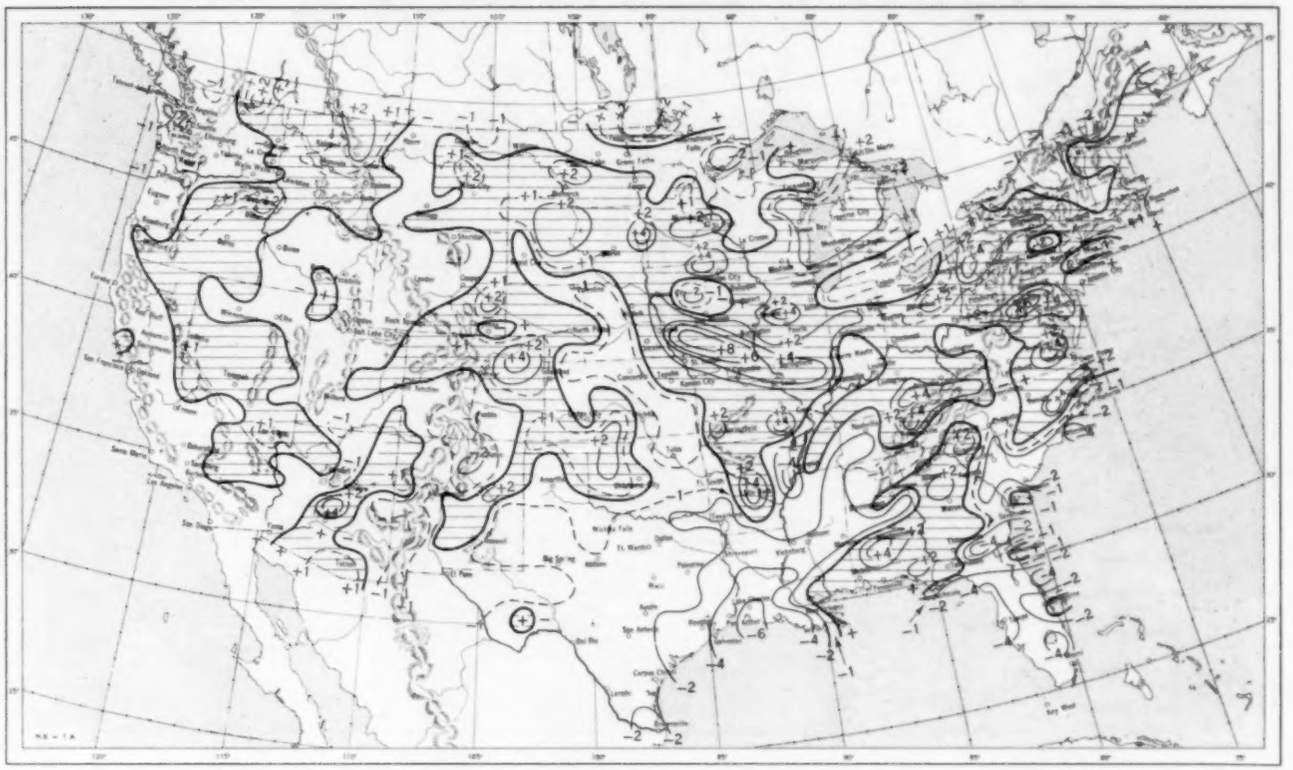
A. Based on reports from 800 Weather Bureau and cooperative stations. The monthly average is half the sum of the monthly average maximum and monthly average minimum, which are the average of the daily maxima and daily minima, respectively.
B. Normal average monthly temperatures are computed for Weather Bureau stations having at least 10 years of record.

Chart II. Total Precipitation (Inches), July 1956.

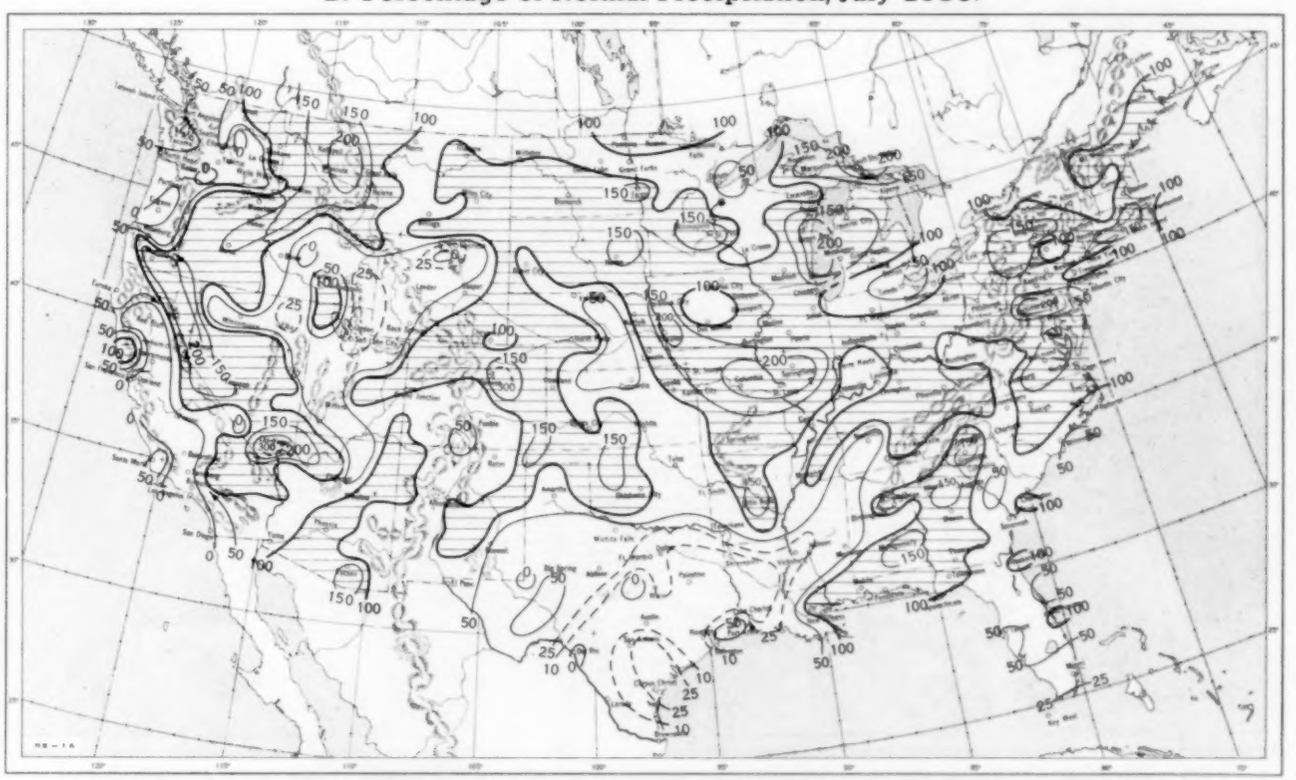

Scale of Shades
0 to 1 inch
1 to 2 inches
2 to 4 inches
4 to 6 inches
Over 6 inches
SCALE OF MILES
LAMBERT CONFORMAL PROJECTION TRUE AT 30° AND 60°

Based on daily precipitation records at 800 Weather Bureau and cooperative stations.

Chart III. A. Departure of Precipitation from Normal (Inches), July 1956.

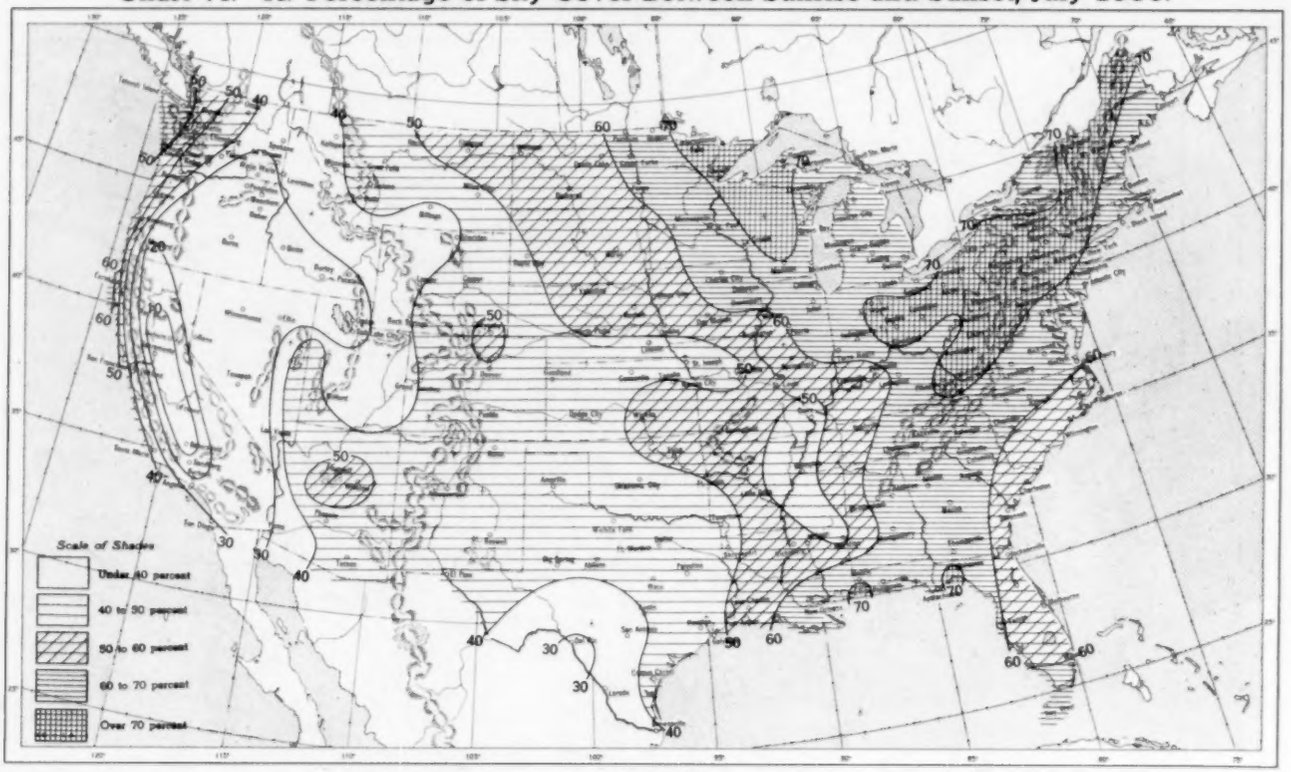


B. Percentage of Normal Precipitation, July 1956.

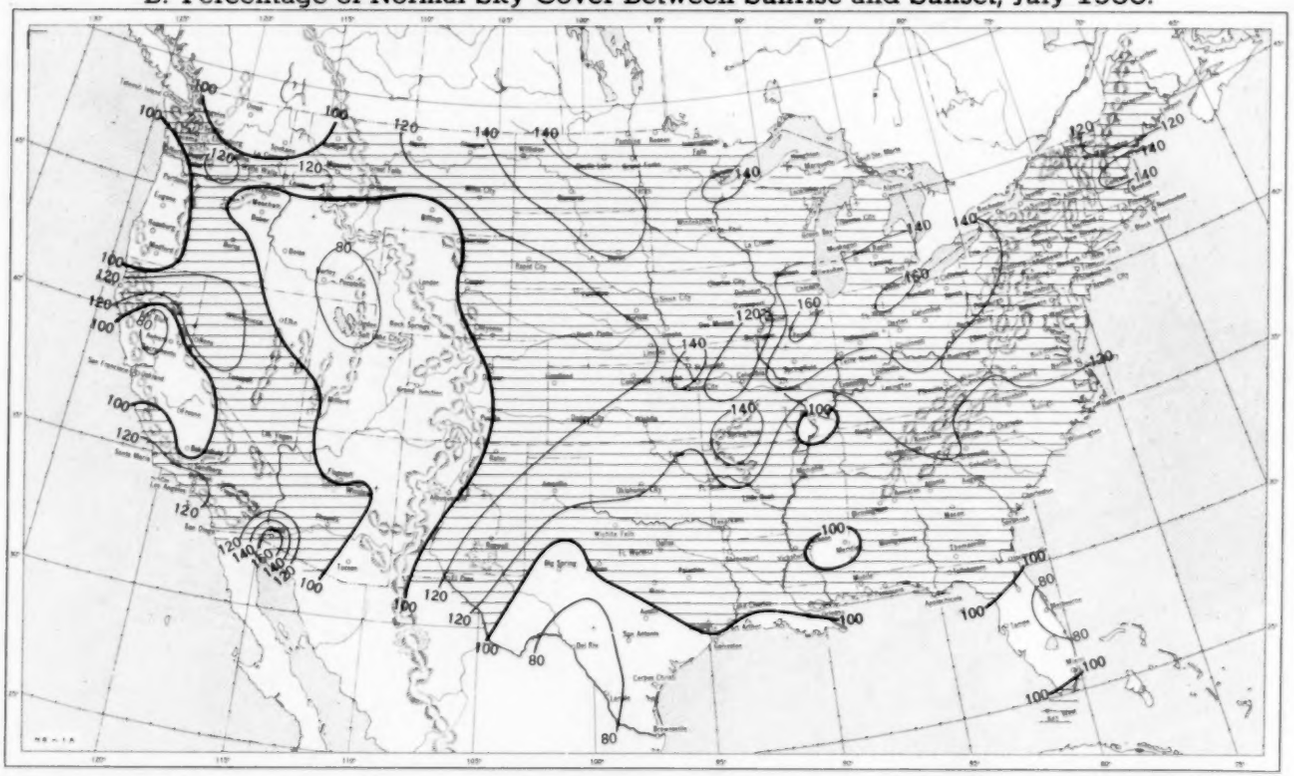


Normal monthly precipitation amounts are computed for stations having at least 10 years of record.

Chart VI. A. Percentage of Sky Cover Between Sunrise and Sunset, July 1956.

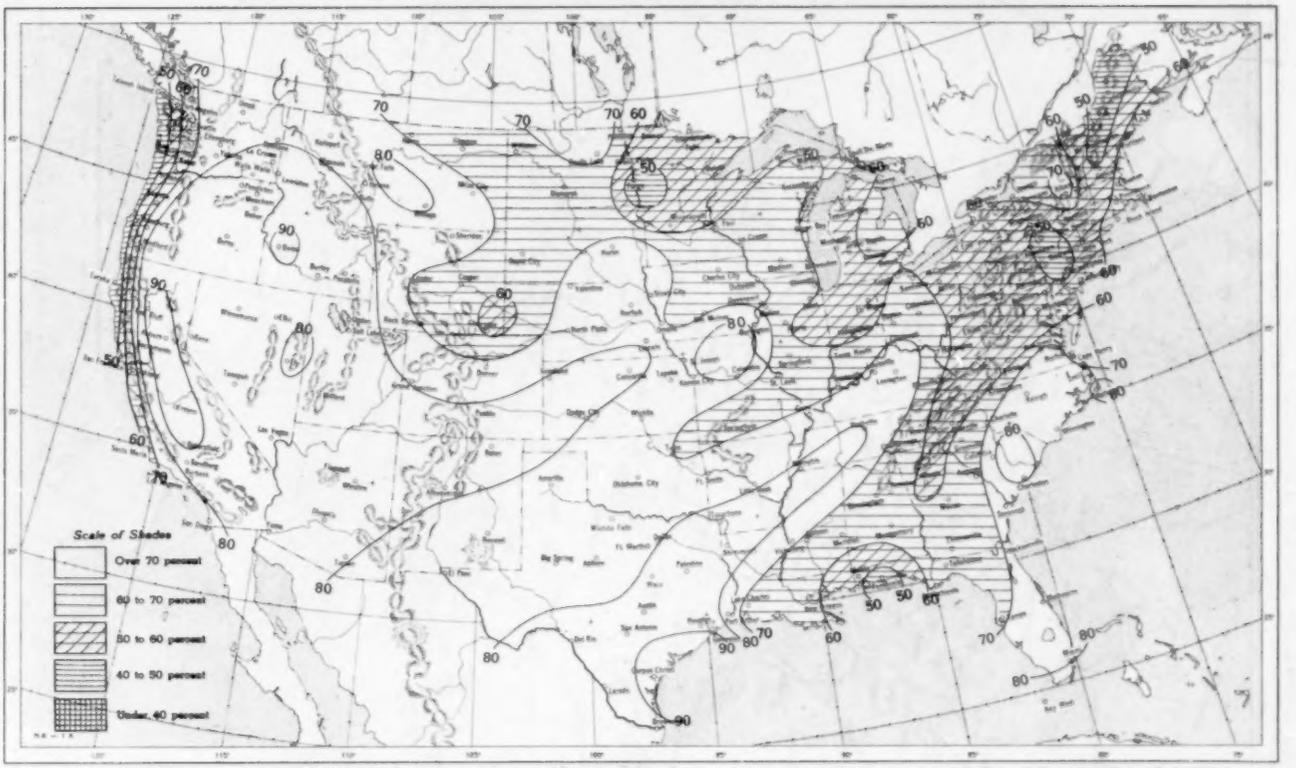


B. Percentage of Normal Sky Cover Between Sunrise and Sunset, July 1956.

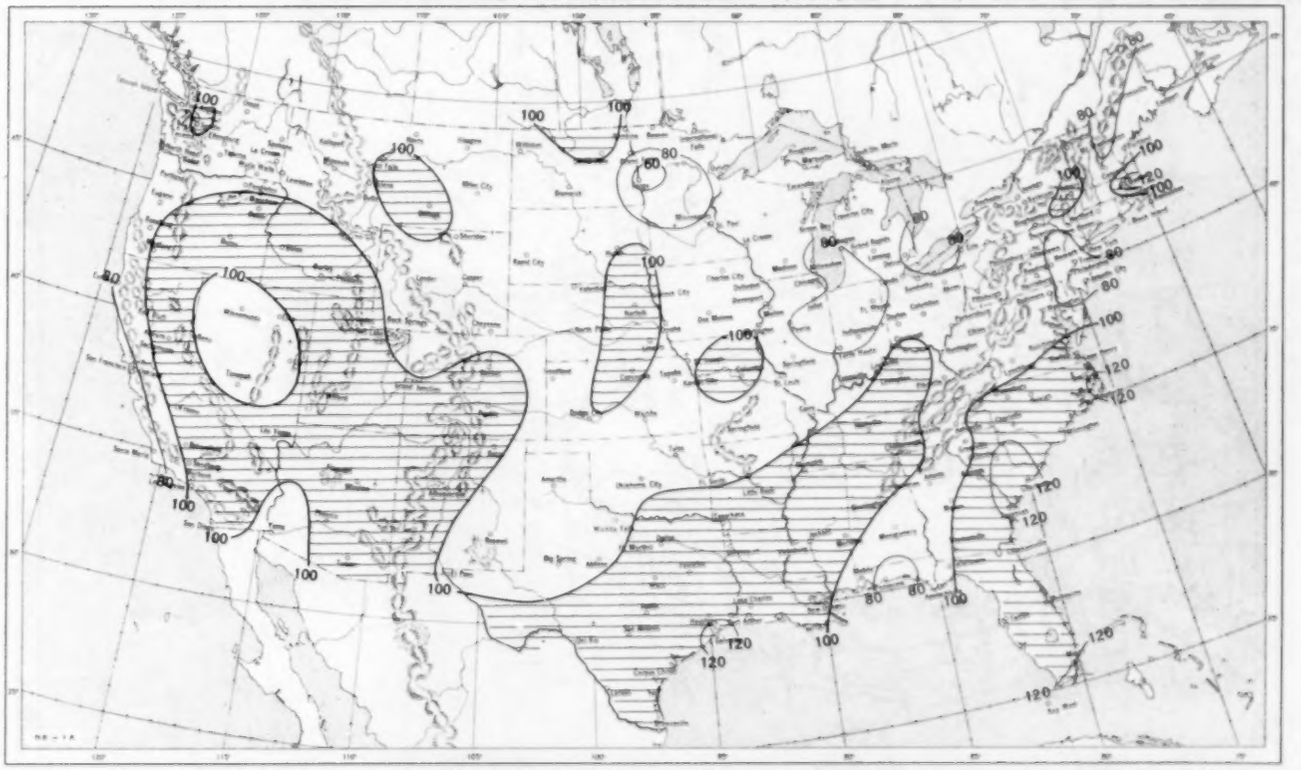


A. In addition to cloudiness, sky cover includes obscuration of the sky by fog, smoke, snow, etc. Chart based on visual observations made hourly at Weather Bureau stations and averaged over the month. B. Computations of normal amount of sky cover are made for stations having at least 10 years of record.

Chart VII. A. Percentage of Possible Sunshine, July 1956.



B. Percentage of Normal Sunshine, July 1956.



A. Computed from total number of hours of observed sunshine in relation to total number of possible hours of sunshine during month. B. Normals are computed for stations having at least 10 years of record.

Chart VIII. Average Daily Values of Solar Radiation, Direct + Diffuse, July 1956. Inset: Percentage of Mean Daily Solar Radiation, July 1956. (Mean based on period 1951-55.)

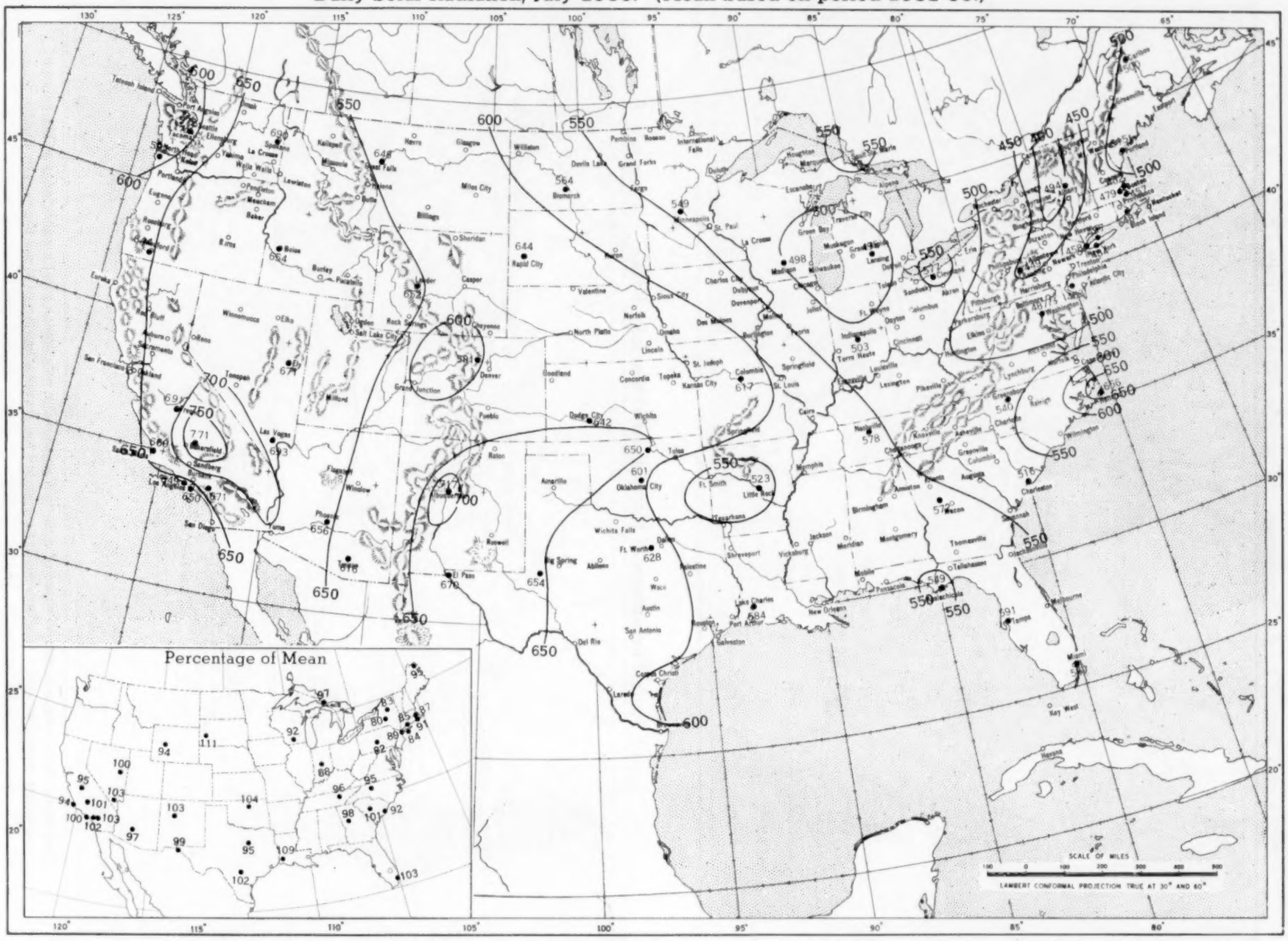
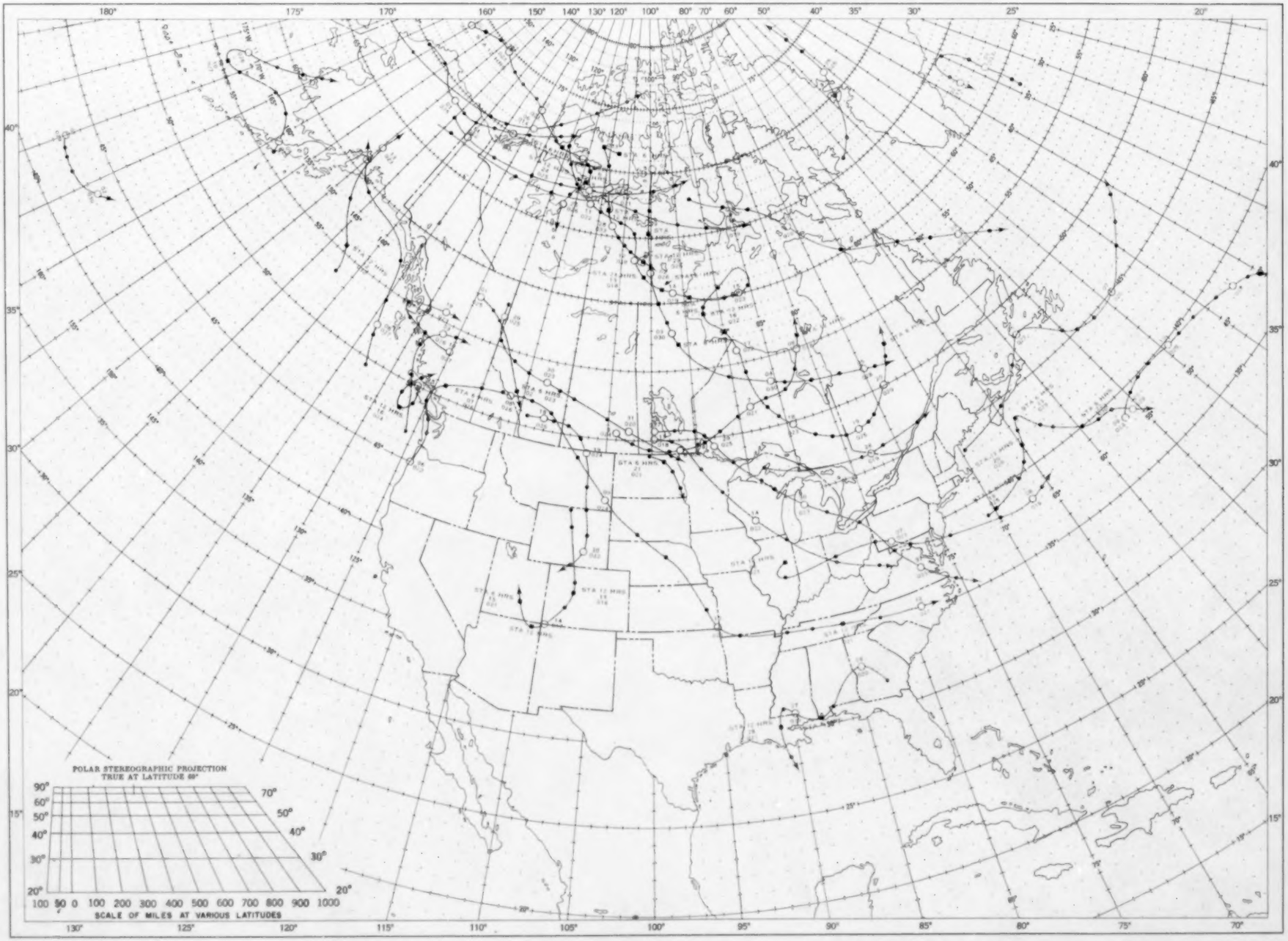


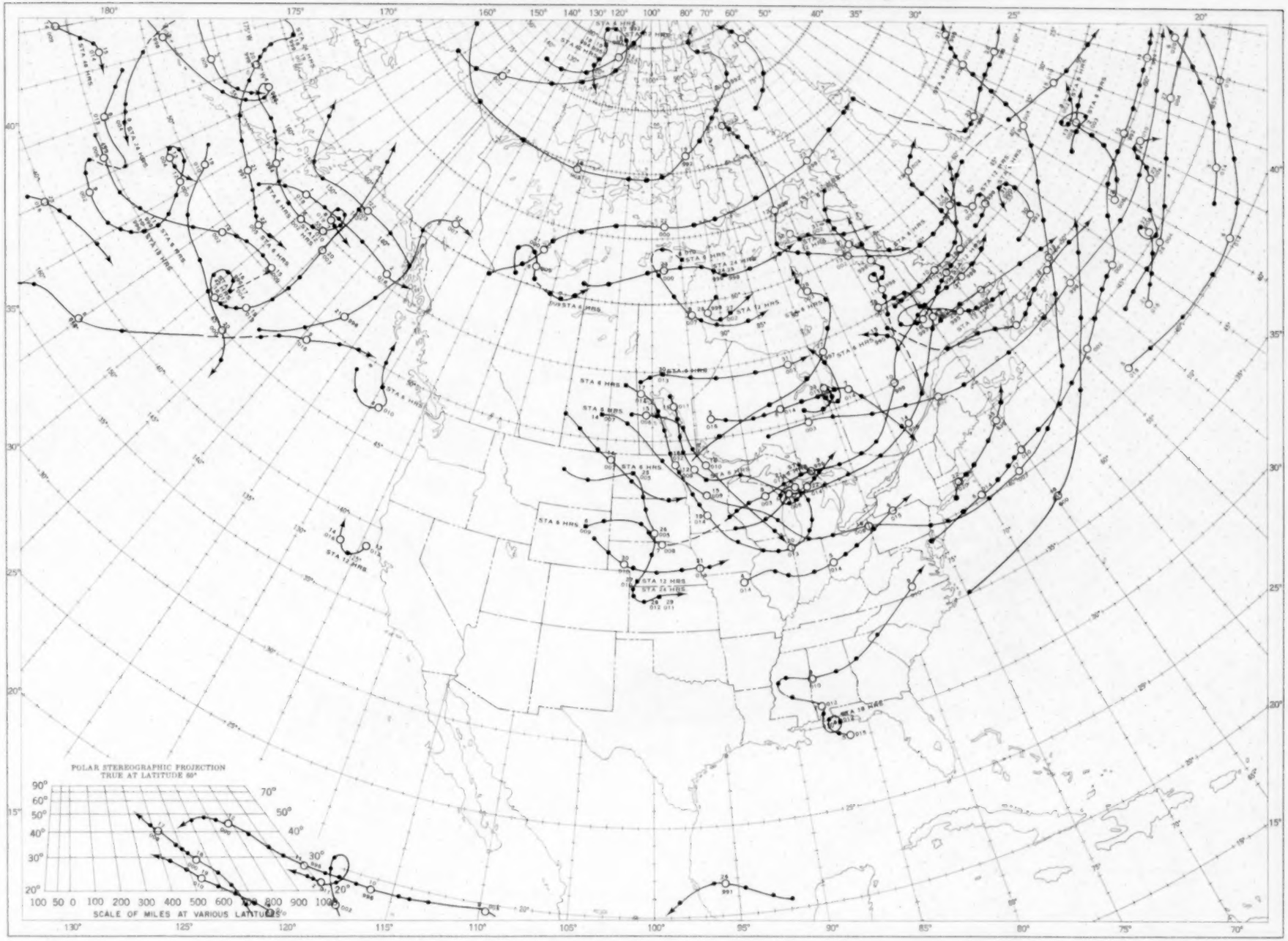
Chart shows mean daily solar radiation, direct + diffuse, received on a horizontal surface in langleys (1 langley = 1 gm. cal. cm. $^{-2}$). Basic data for isolines are shown on chart. Further estimates are obtained from supplementary data for which limits of accuracy are wider than for those data shown.

Chart IX. Tracks of Centers of Anticyclones at Sea Level, July 1956.


POLAR STEREOGRAPHIC PROJECTION
TRUE AT LATITUDE 60°
SCALE OF MILES AT VARIOUS LATITUDES
100 50 0 100 200 300 400 500 600 700 800 900 1000

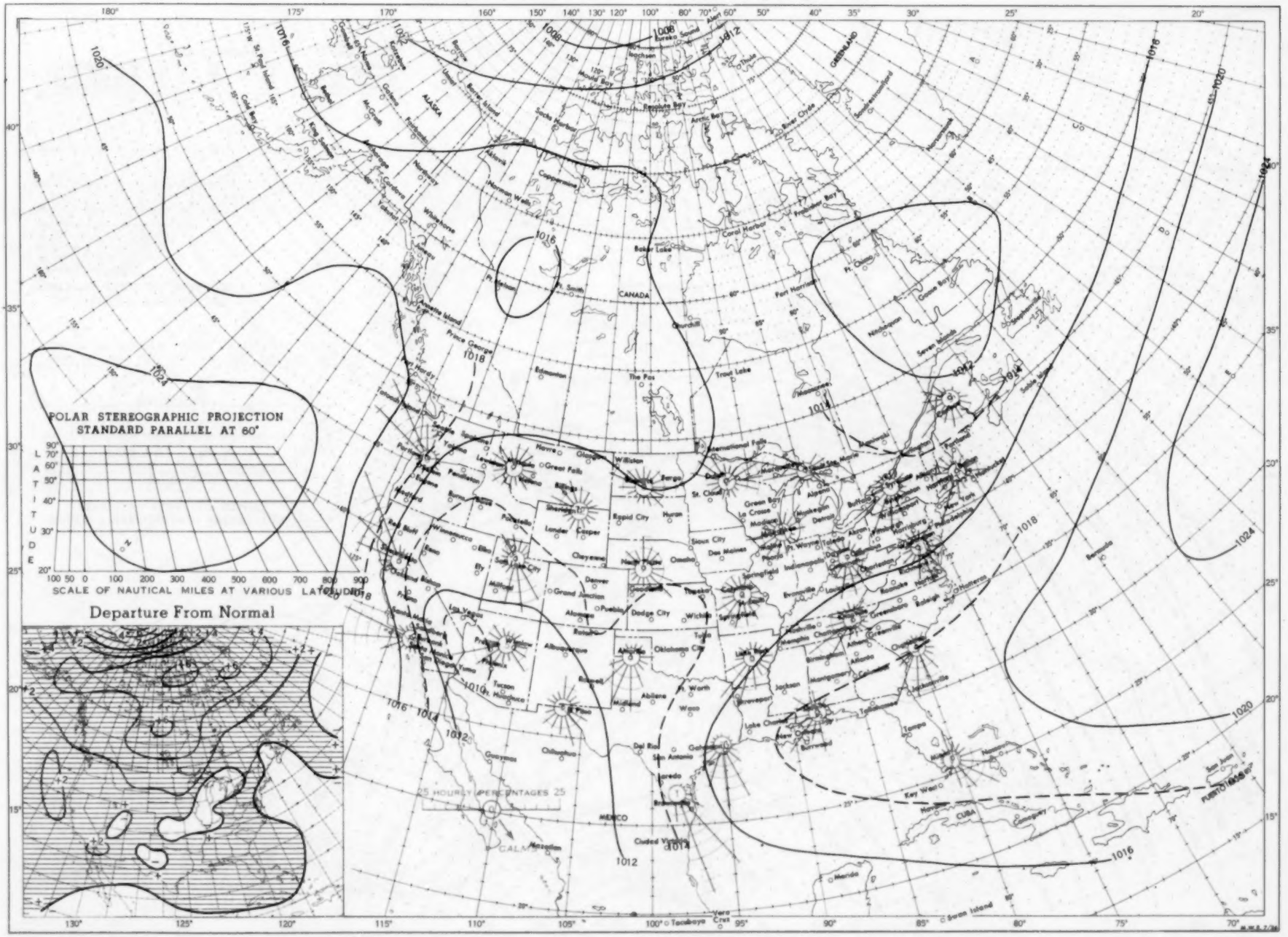
Circle indicates position of center at 7:30 a. m. E. S. T. Figure above circle indicates date, figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Squares indicate position of stationary center for period shown. Dashed line in track indicates reformation at new position. Only those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, July 1956.



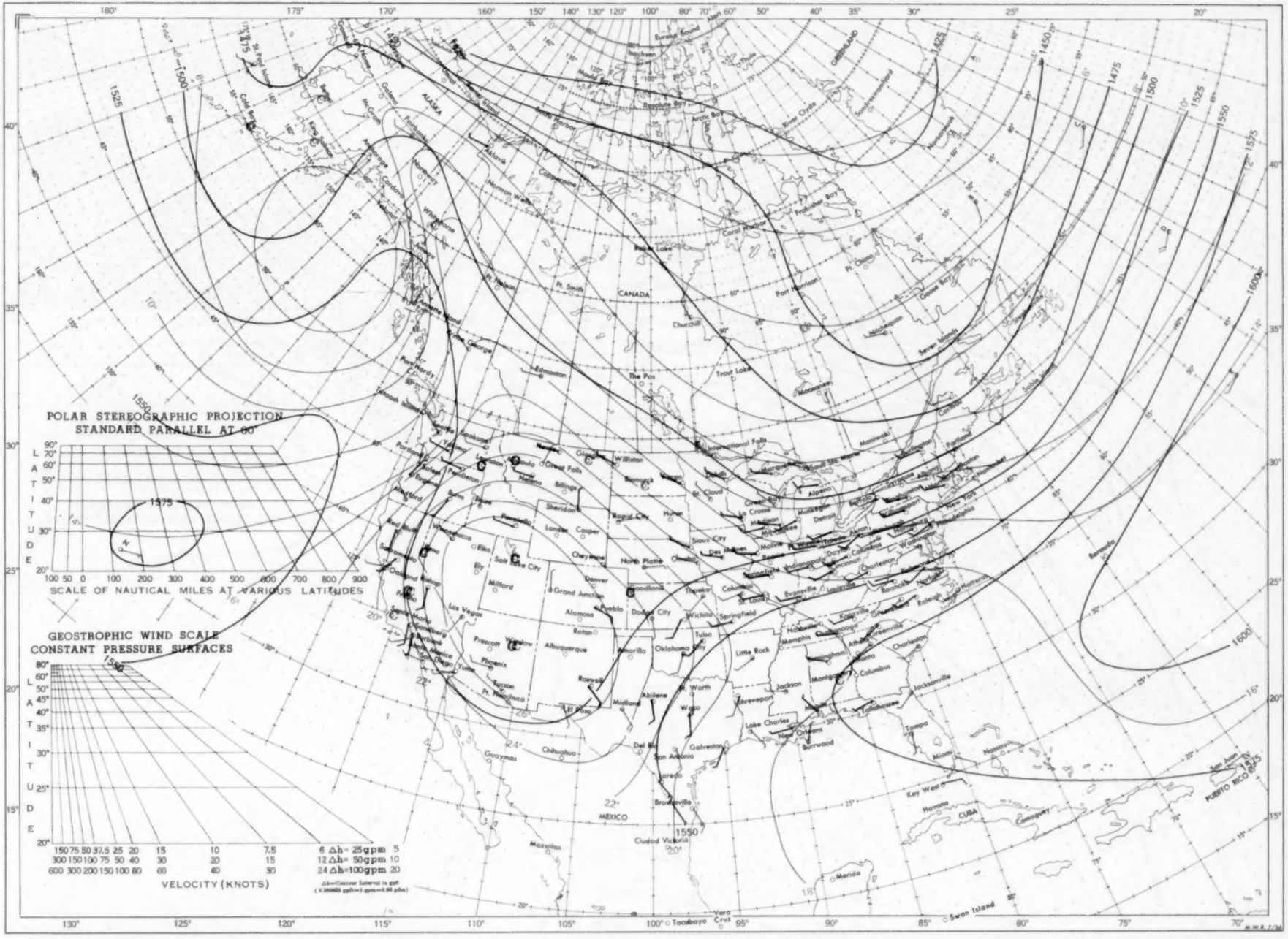
Circle indicates position of center at 7:30 a. m. E. S. T. See Chart IX for explanation of symbols.

Chart XI. Average Sea Level Pressure (mb.) and Surface Windroses, July 1956. Inset: Departure of Average Pressure (mb.) from Normal, July 1956.



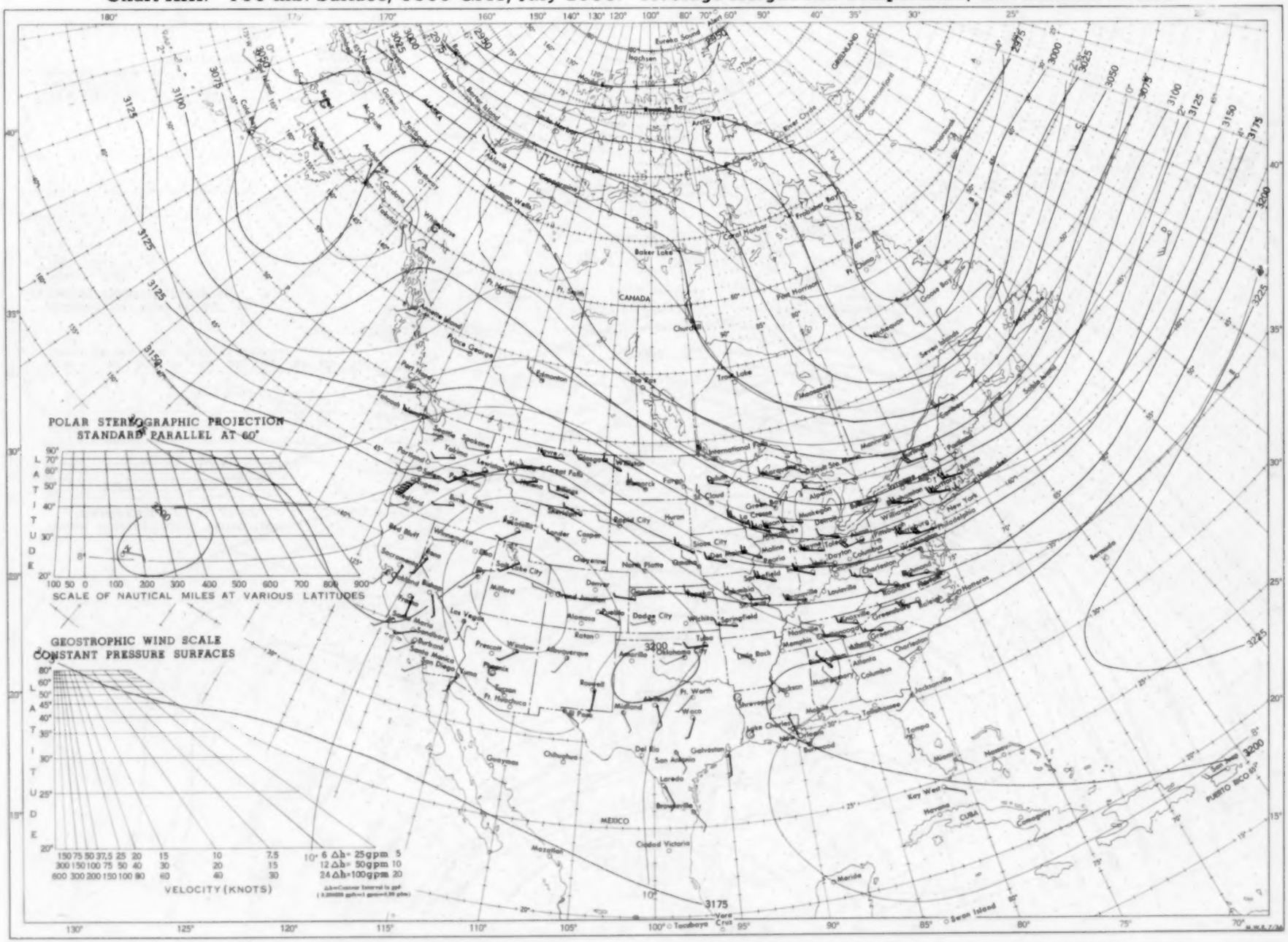
Average sea level pressures are obtained from the averages of the 7:30 a.m. and 7:30 p.m. E.S.T. readings. Windroses show percentage of time wind blew from 16 compass points or was calm during the month. Pressure normals are computed for stations having at least 10 years of record and for 10° intersections in a diamond grid based on readings from the Historical Weather Maps (1899-1939) for the 20 years of most complete data coverage prior to 1940.

Chart XII. 850-mb. Surface, 0300 GMT, July 1956. Average Height and Temperature, and Resultant Winds.



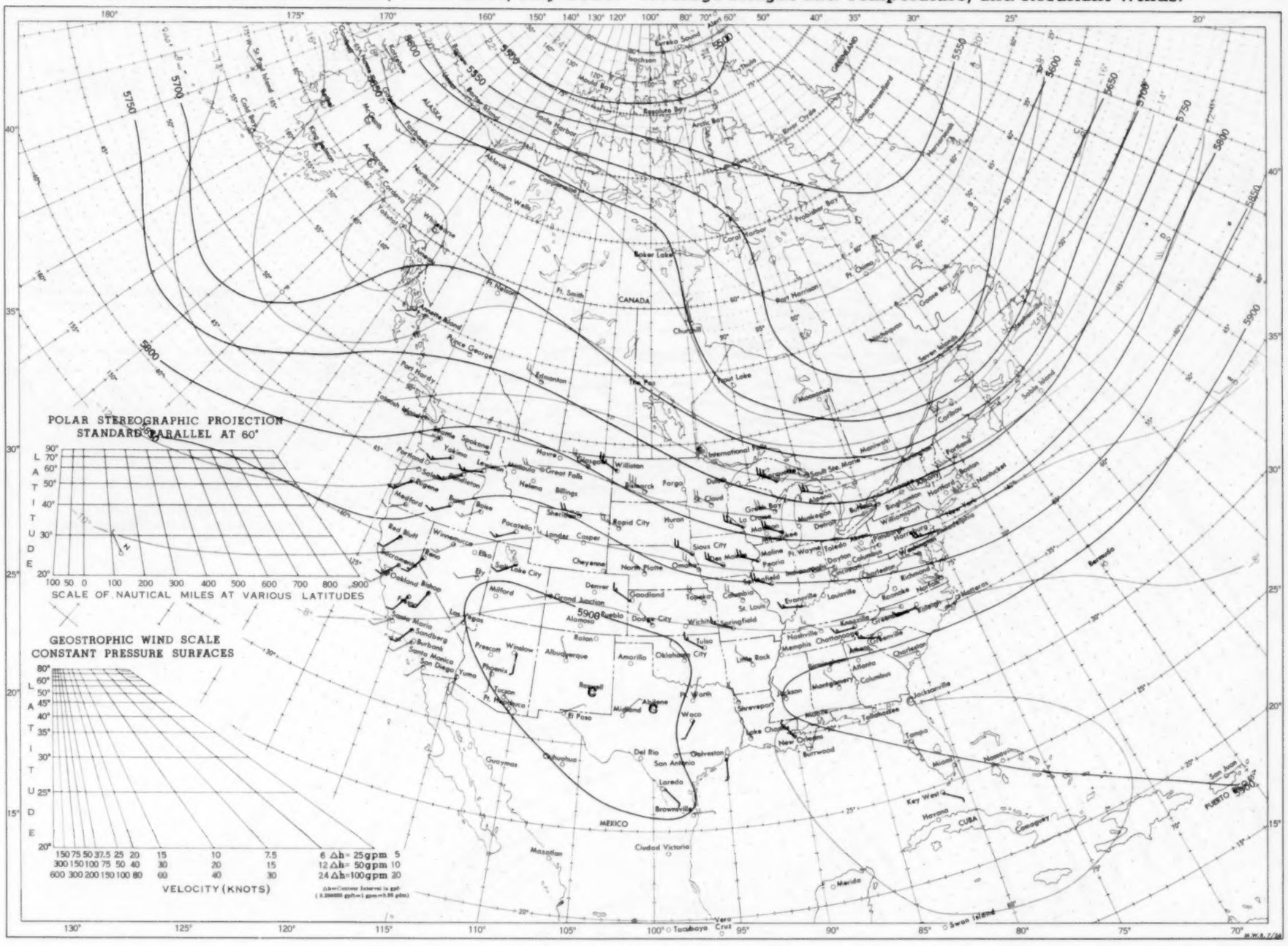
Height in geopotential meters (1 g.p.m. = 0.98 dynamic meters). Temperature in °C. Wind speed in knots; flag represents 50 knots, full feather 10 knots, and half feather 5 knots. Winds shown in red are based on rawins taken at the indicated pressure surface and time. Those in black are based on pibals taken at 2100 GMT and are for the nearest standard height level.

Chart XIII. 700-mb. Surface, 0300 GMT, July 1956. Average Height and Temperature, and Resultant Winds.



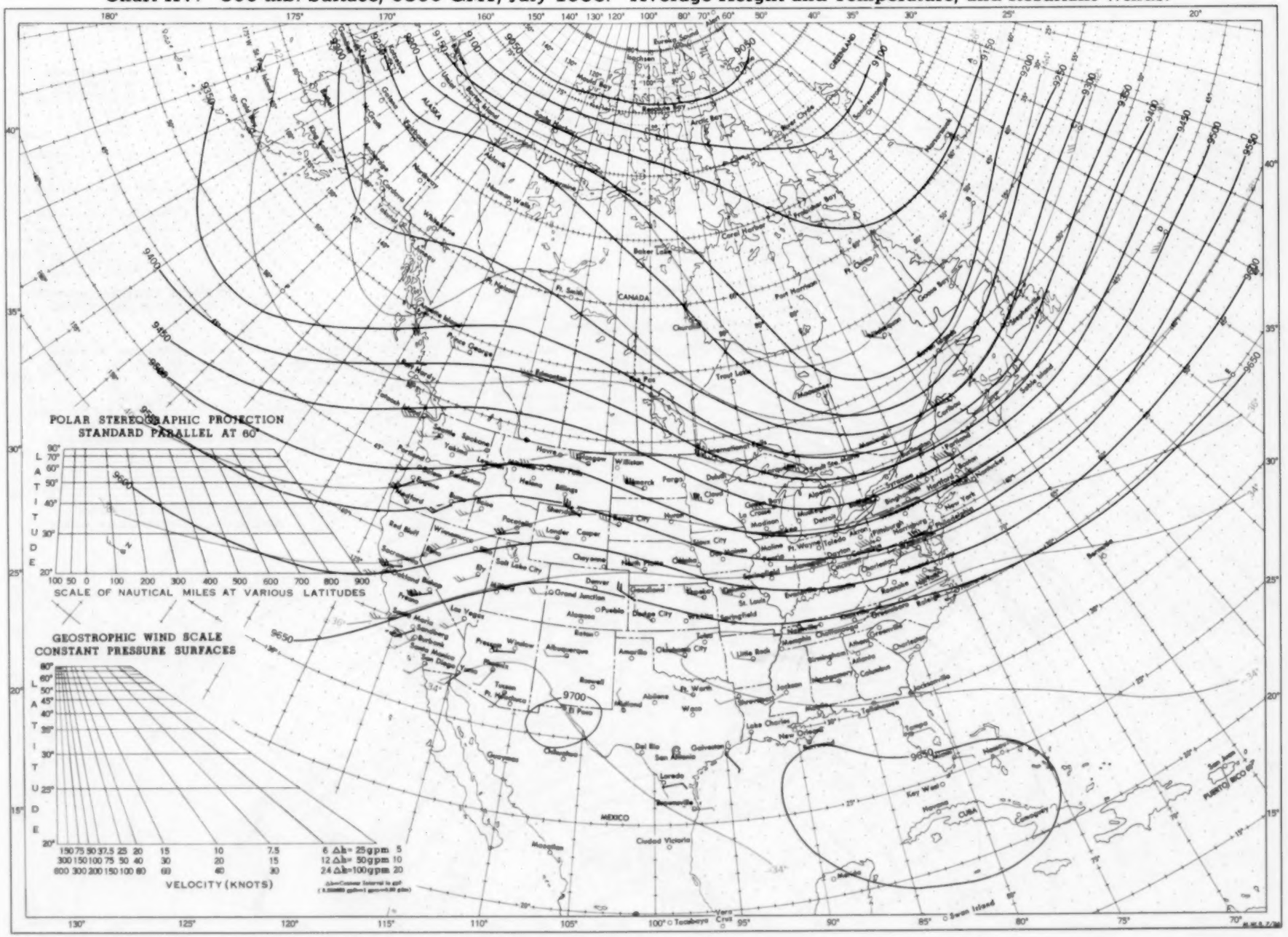
See Chart XII for explanation of map.

Chart XIV. 500-mb. Surface, 0300 GMT, July 1956. Average Height and Temperature, and Resultant Winds.



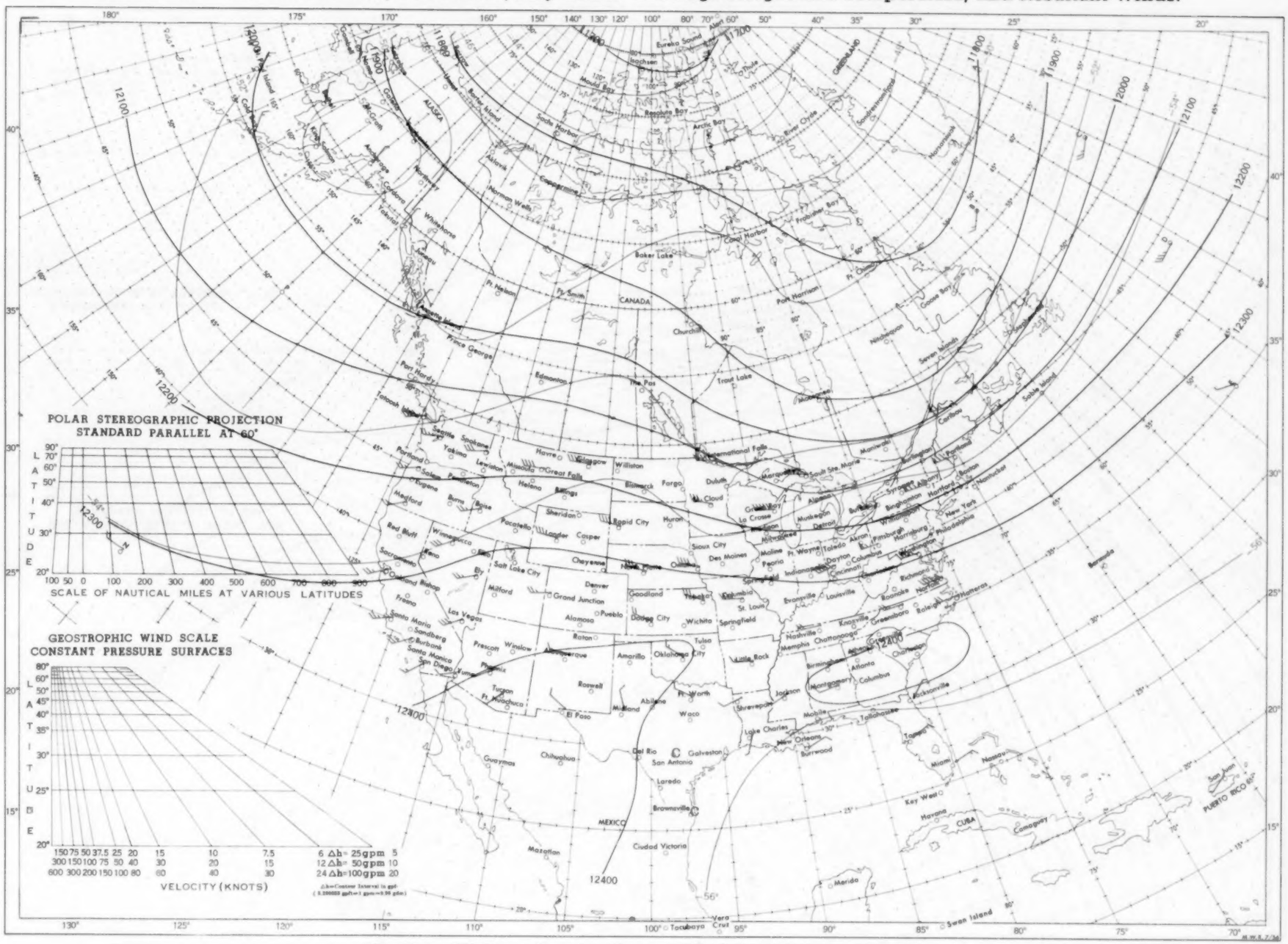
See Chart XII for explanation of map.

Chart XV. 300-mb. Surface, 0300 GMT, July 1956. Average Height and Temperature, and Resultant Winds.



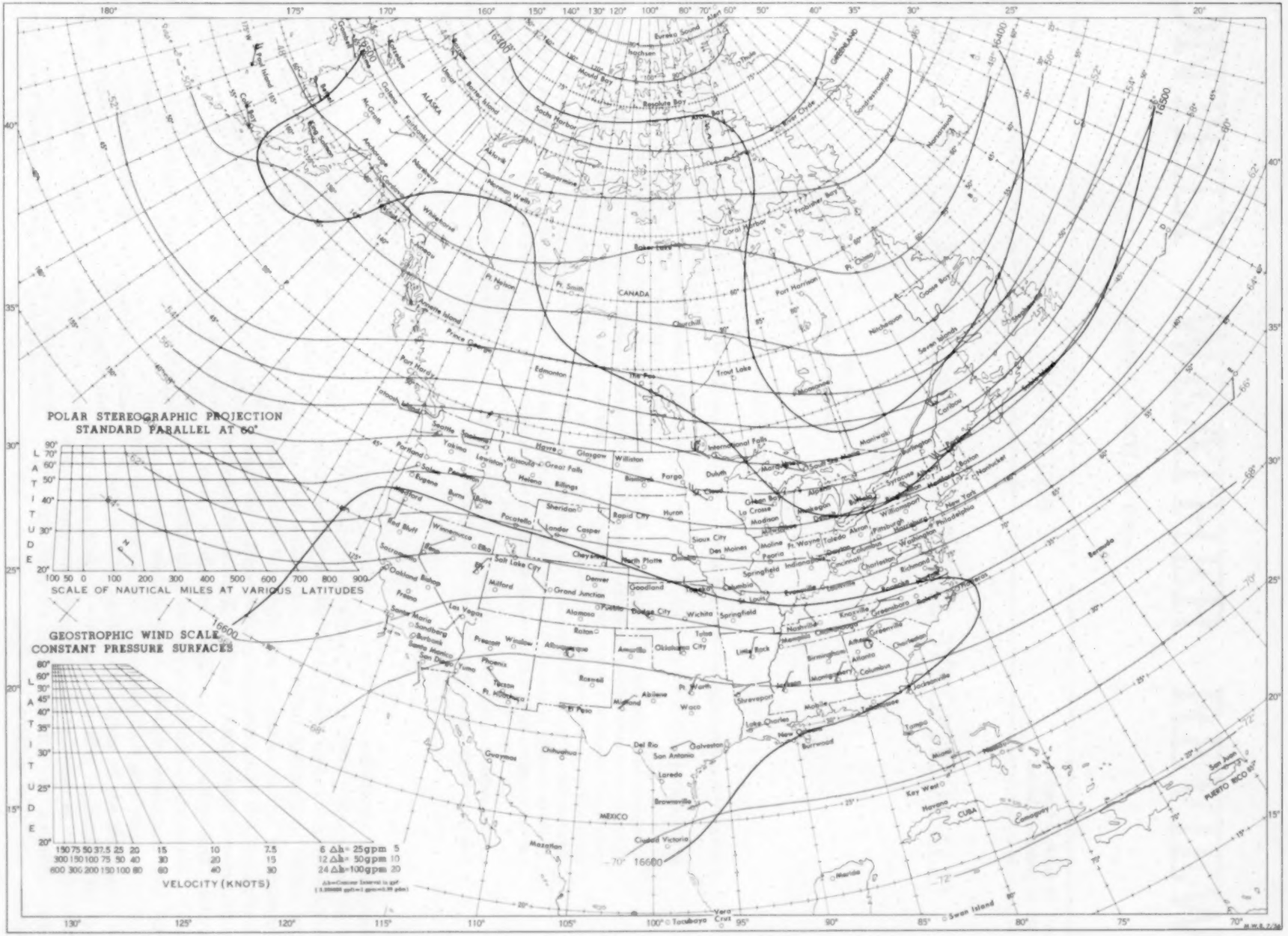
See Chart XII for explanation of map.

Chart XVI. 200-mb. Surface, 0300 GMT, July 1956. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map. All winds are from rawin reports.

Chart XVII. 100-mb. Surface, 0300 GMT, July 1956. Average Height and Temperature, and Resultant Winds.



See Chart XII for explanation of map. All winds are from rawin reports.